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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 40c.)

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FEBRUARY 1958

ORDOVICIAN HIGHGATE SPRINGS SEQUENCE OF VERMONT AND QUEBEC AND ORDOVICIAN CLASSIFICATION

MARSHALL KAY

ABSTRACT. The Highgate Springs slice, eighty miles long and a mile or so wide, extends from north of St. Dominique, Bagot County, Quebec, to northwestern Chittenden County, Vermont. The Canadian Beldens limestone persistently forms the base of the stratigraphic sequence. In the Chazyan, argillaceous quartz sandstone is succeeded by varying proportions of limestone, argillaceous limestone and calcareous argillite—the Carman quartzite and Youngman formation in Vermont and extreme southern Quebec, and the St. Dominique formation in Bagot County, Quebec. The Blackriveran Isle la Motte limestone, a constant unit of a few tens of feet, is succeeded by the Trentonian Glens Falls limestone and Hortonville shale. The slice lies on the Highgate Springs thrust fault, which crosses the structures in the Ordovician shales beneath and to the west; the slice is cut off on the east by the Granby thrust, Philipsburg thrust and Rosenberg thrust beneath slices of dissimilar rocks.

Of greatest interest stratigraphically is the Chazyan Series, which contains a sparse fauna that in the argillaceous facies contains fossils not known from the type Chazyan and supposed to be distinctive of post-Chazyan rocks; thus the Highgate Springs sequence bears on the classification of the stages in the Ordovician System. Structurally, the slice seems to be the attenuated east limb of the essentially synclinorial structure of the northern Champlain Lowland.

INTRODUCTION

The Ordovician rocks at Highgate Springs, Franklin County, northern Vermont, are of a distinctive sequence that outcrops in scattered areas from east of St. Hyacinthe, Quebec, to southwest of St. Albans, Vermont, for about 80 miles (fig. 1). The succession has been called the Highgate Springs Series (McGerrigle, 1930, p. 131), the lower part being Division E of the Quebec Group of Sir William Logan (1863, p. 854); the term Highgate Springs sequence is here used to avoid the connotation that it is a formal stratigraphic series. The sequence is discussed as forming a single slice, though there are faults and folds within it.

The sequence lies on a thrust fault, the Highgate Springs thrust, separating it from the relatively gently deformed rocks of the structural foreland (Hawley, 1957) along Lake Champlain and the upper Richelieu River. The principal structure (Clark, 1934; 1955) is a northeast-pitching synclinorium having an axial trend somewhat more northeasterly than the fault trace (Hawley, 1957). On the east, from Swanton northward into Quebec, the Philipsburg thrust separates the Philipsburg sequence, a northeast-pitching syncline of Ordovician (Clark, 1941) from the underlying Highgate Springs slice, or the foreland rocks where that is cut out along the fault trace. The Rosenberg thrust lies east of the Philipsburg sequence and, as the Champlain thrust, adjoins the Highgate Springs slice from Swanton southward, coming to adjoin the foreland from Malletts Bay southward. The sole of the Rosenberg

slice is Lower Cambrian dolomite. In west-central Vermont, the Middlebury Synclinorium, plunging south and lying east of the Champlain thrust, has a core of Ordovician sediments most like those in the Highgate Springs slice (Cady, 1945).

This very narrow belt of rocks of such great linear extent, the westermost belt of strong deformation, provides fuller understanding of the structure and structural history of western New England.

The Highgate Springs sequence consists of diverse sediments ranging from uppermost Canadian to upper Trentonian. Equivalent rocks are well

Fig. 1. Map of the outcrop belt of the Highgate Springs sequence in Quebec and Vermont. Rectangles indicate areas of figures 6, 5 and 7 (north to south).

shown in the foreland sequence, and some of them are quite alike. The Philipsburg sequence, which lies to the east, encompasses this time span, but there is little obvious similarity. The Rosenberg slice farther east is of older rocks, but its structural continuation to the south, the Middlebury Synclinorium, has formations resembling those in the Highgate Springs sequence.

The Highgate Springs sequence has received only brief and cursory attention since the study by Sir William Logan a century ago. There is more than local interest in the structure and stratigraphy. The rocks of the sequence are exposed interruptedly for about 80 miles along a belt that is generally only a few hundred yards broad, rarely a mile. The intricate local structure of this peculiarly narrow slice is challenging to resolve. But the structural setting is more difficult to understand—that a few hundred feet of section of constant stratigraphic span should persist in between older rocks of several thrust slices on the east, and younger argillites on the west. Some of the rock types in the sequence change significantly along the trend, and contrast with rocks of the same age on the west. Knowledge of the sequence adds measurably to understanding the paleogeography as well as the structural history of medial Ordovician time.

The field studies were carried forward interruptedly during the past twenty years, partly supported by Graduate Funds of Columbia University. Graduate students made large-scale plane table maps of the areas of largest exposure at Highgate Springs and Fonda Junction. Studies of the Ordovician rocks in surrounding regions by T. H. Clark, W. M. Cady, Philip Oxley and L. D. Hawley have increased our understanding of the significance of local observations. John Chatfield prepared most of the illustrations. G. A. Cooper kindly aided in the identification of brachiopods.

This text presents first a summary of the geology of the Highgate Springs area, and then of the local outcrop areas to the north in Quebec and south in Vermont. There follows a discussion of the stratigraphic units and their regional stratigraphy, then of the structural significance of the slice.

LOCAL GEOLOGY

The Highgate Springs area.—The geology around Highgate Springs was crudely sketched by Hitchcock (1861, p. 283), and the principal lithic units and structures were outlined by Logan (1863, p. 273-275, 855-856), Brainerd (1891, p. 298) Gordon (1923) and the writer (Kay, 1937, p. 254) for the upper part of the section. The stratigraphy and structure have not been adequately described and mapped.

The rocks are exposed in a number of linear ridges separated by meadows. They are the limbs of asymetrical folds with east-dipping axial planes, broken by low-angle faults that produce interruptions and repetitions, and by a few offsetting high-angle faults that cross the dominant trends. The geologic map (fig. 2) shows the principal outcrop area, though similar structure continues half mile to the south to Youngman Creek, and there are good exposures on Limekiln Point, across Phelps Bay on the north. The most obvious rock is a massive limestone; it forms the long U-shaped band of ledges that opens northward from the roadside west of the center of the village and

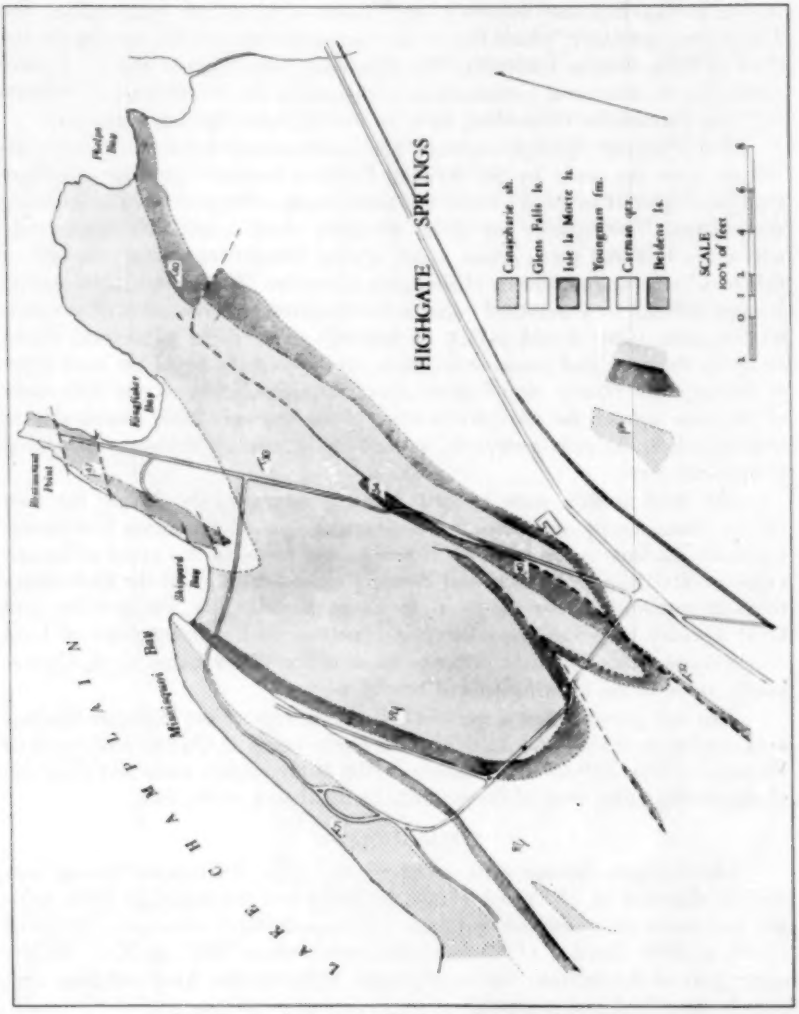

Fig. 2. Geologic map, Highgate Springs, Vermont.

delineates a south-pitching anticline. Progressively older beds within the core of the anticline are argillaceous limestones, quartz sandstones, and interbedded white-weathering calcite limestones and buff-weathering dolomites that extend on to Restaurant Point and the island to the north. The units succeeding the massive, ledge-forming limestone are dark, thin-bedded fossiliferous argillaceous limestones and dark shaly argillites. The six divisions recognized by Logan as described by him (1863, p. 273-274) but with present names and recently determined thicknesses, are from youngest to oldest:

"Black, brittle shales" "of the Utica Formation"—the Hortonville shale exposed for about 20 feet

"Black thin bedded shaly nodular limestones" that "belong to the Trenton" Glen Falls limestone, about 100 feet

"Massive black limestone" the Isle la Motte limestone, about 70 feet

"Blackish thin bedded shaly nodular limestones", the Youngman formation, only partly exposed, some 300 feet

"Greenish-grey calcareous fine grained sandstones" . . . the Carman sandstone, some 120 feet

"Dove-grey limestone . . . associated with bands of brownish-grey, buff-weathering dolomite", the Beldens formation, exposed for about 150 feet

Logan did not recognize the complexity of the structure, and hence misjudged thicknesses, giving 50 to 100 feet, more than 60 feet, about 30 feet and between 300 and 400 feet for the Carman, Youngman, Isle la Motte, and Glens Falls respectively.

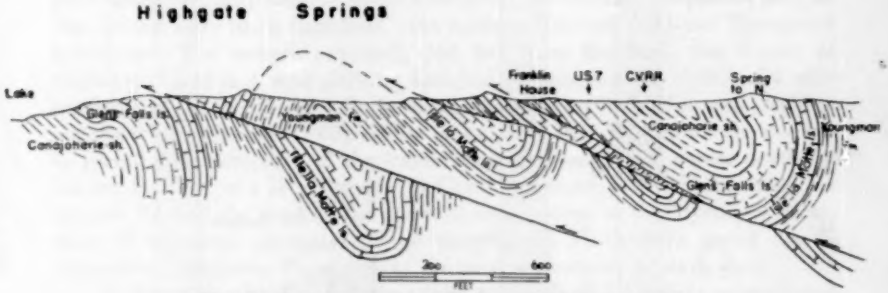

Fig. 3. Structure section, Highgate Springs, Vermont.

The Highgate Springs Sequence thus consists of a few hundred feet of Ordovician limestones, sandstones and shales, with Beldens at the base and the black "Utica" shale at the top (fig. 4). It is unlike the nearest rocks either to the east or to the west, which are classified in other sequences. The nearest exposures to the east, at an abandoned limekiln a quarter mile from the village, consist of a belt of white calcitite having patches of dolomitization, a "curdled" structure, and a few beds with rounded quartz sand grains. This rock has been thought to be in the Upper Cambrian Rock River formation in the Philipsburg Sequence (McGerrigle, 1930), a formation that the Philipsburg thrust throws against the Ordovician black shale at the shore of Missisquoi Bay from Philipsburg to the International Boundary.

About a mile east of Highgate Springs is the continuous escarpment of Lower Cambrian dolomite of the Rosenberg sequence, overlying the Rosenberg or Champlain thrust; and Cambrian rocks continue eastward into the Green Mountains. West of Highgate Springs is the delta of the Missisquoi River; there are scattered exposures of black shale on Hog Island, a peninsula beyond. Rocks like those in the Highgate Springs Sequence next appear in the foreland on Isle la Motte in Lake Champlain ten miles farther westward.

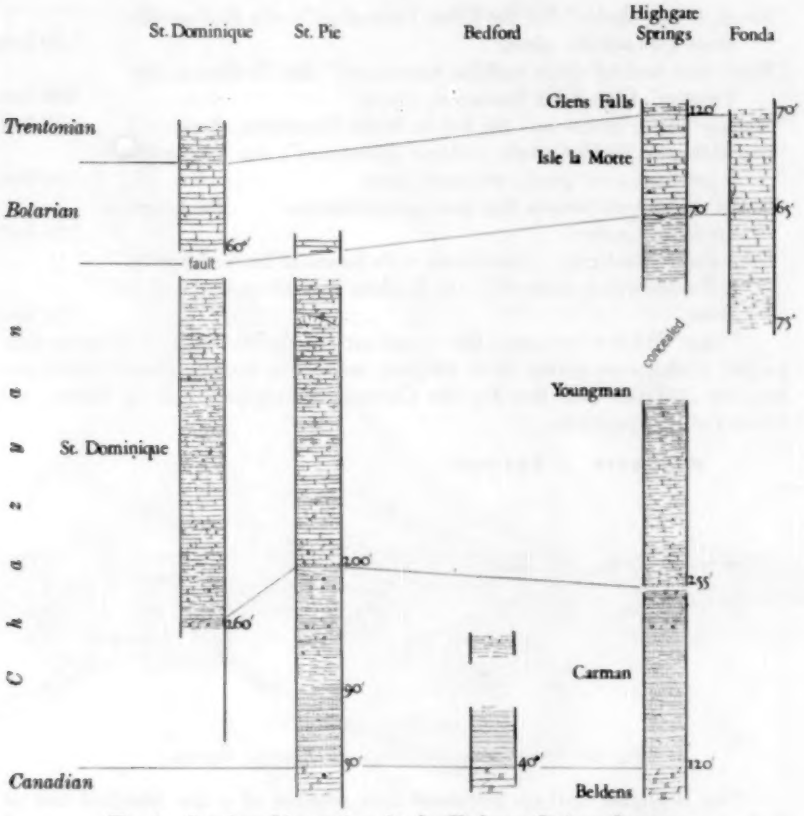

Fig. 4. Stratigraphic sections in the Highgate Springs Sequence.

There, Isle la Motte limestone underlies Glens Falls limestone, much as at Highgate Springs, but lies on Chazy limestone unlike rocks in the typical Highgate Springs sequence (Oxley, 1952); a basal quartzite resembling the Carman quartzite lies on Beekmantown Bridport dolomite like the dolomite interbeds in the Beldens, but lacking the dominant laminated limestone.

On the other hand, rocks like those at Highgate Springs are found interruptedly to the north in a continuing trend as far as St. Dominique, Bagot County, Quebec, some fifty miles away, and southward along the Lake Champlain shore to Camp Rich, twenty miles away in northernmost Chittenden County, Vermont, see figures 6 and 7.

AREAS OF HIGHGATE SPRINGS SEQUENCE IN QUEBEC

Rocks similar to those at Highgate Springs are exposed in but two areas northward in Quebec, a very small one in Stanbridge Township near Bedford, Missisquoi County and a much larger one extending from north of St. Pie to northwest of St. Dominique de Bagot in Bagot County.

Fig. 5. Map of exposures of Highgate Springs Sequence in Stanbridge Township, near Bedford, Quebec.

Stanbridge area.—Two isolated exposures lie in lot 5, Conc. 8, Stanbridge Township, east of the Central Vermont Railroad south of the road crossing about a mile southeast of Stanbridge Station (fig. 5). The rocks, referred to "division E" by Logan (1863, p. 854), are "seventy feet of dove-grey limestone, fifty feet of white-weathering sandstone, and twenty feet of thin-bedded slaty black limestone", the Beldens, Carman and basal Youngman formations. The western exposure, 100 feet from the track, has a core of marble enclosed in a west-pitching anticline of quartzite. An eighth of a mile southeast is a ridge of quartzite forming the northeast limb of an east-pitching anticline with dips of 30 degrees or so on the north. Exposures are too few to permit reconstruction of the intervening structure. Half a mile northeast toward Bedford is a large quarry in Corey limestone of the Philipsburg Sequence. Beyond the road to the north is an exposure of the Mystic conglomerate of the same sequence, so the trace of the Philipsburg thrust lies in intervening meadows. Exposures to the west are entirely of dark slate.

St. Pie Hill.—St. Pie Hill rises from the lowland 1.5 miles east-northeast of St. Pie, Bagot County. It forms the southern end of an escarpment of discontinuous exposures extending for eight miles northward to beyond St. Dominique. The best exposures are at St. Pie Hill and four miles northward along Highway 12 west of St. Dominique; the areas are discussed separately. The section has been noted by Clark (1955, p. 11).

The principal exposures on St. Pie Hill are in a homocline dipping 35 to 45 degrees east (fig. 6). On the west is a cliffed escarpment of quartzite and argillaceous shale, beds that partially surround a south-pitching anticline with a core of white calcitite and buff-weathering dolomite, the Beldens limestone. On the west, these rocks are separated by a thrust fault from exposures of vertical to steeply west-dipping fossiliferous Glens Falls limestone; the fault trends more northerly than the beds to east, so that the Beldens lies against Glens Falls near the road north of the hill, but the limestones in the stratigraphic position of the Youngman adjoin it at the south. At the farm buildings northeast of the hill, east of the farmhouse, dark, aphanitic "Isle la Motte" limestone is poorly exposed and fossiliferous Glens Falls limestone dips steeply to the east.

St. Dominique.—A number of quarries have been opened in the limestones on both sides of the St. Dominique—St. Hyacinthe highway, in the

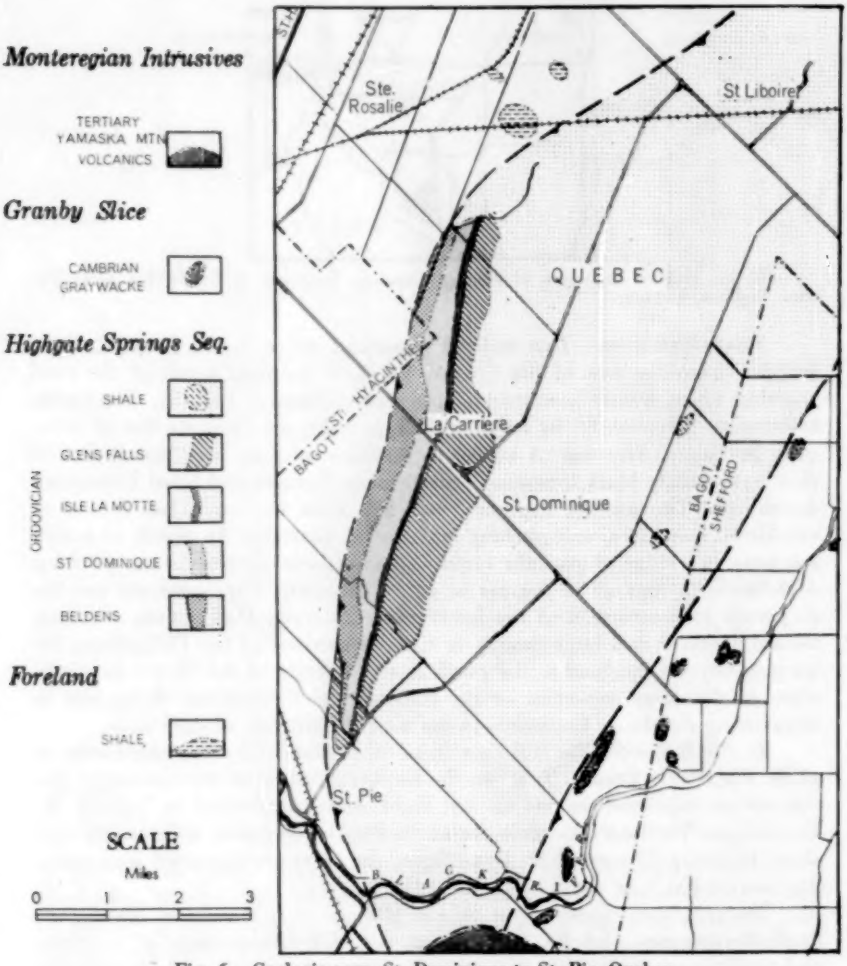

Fig. 6. Geologic map, St. Dominique to St. Pie, Quebec.

escarpment about a mile west of St. Dominique. The rocks in the quarries have been analyzed by Goudge (1935). On the west, Carman quartzite crops out in the lowest continuous exposures, but white-weathering calcitite and buff-weathering dolomite of the Beldens crop out in a quarry and adjoining low hills half a mile north of the highway west of the escarpment. The principal exposures are in large quarries in the siliceous limestone overlying the Carman and having the stratigraphic position of the Youngman; the dips are generally about 15 to 20 degrees southeast to south-southeast.

Beyond the crest of the hill and east of the siliceous limestones is a belt of high-calcium limestone, the "Isle la Motte." The contact of the two forma-

tions is a fault dipping 75 degrees or so southeast. The heavy ledges of relatively pure Isle la Motte, dipping about 20 degrees southeastward, attain an exposed thickness of 40 feet in the La Pointe quarry south of the road, and the belt has been excavated for a mile southward. To the east is fossiliferous thin-bedded Glens Falls limestone and, beyond St. Dominique, the succeeding dark shaly argillite.

St. Hyacinthe is about 4 miles northwest of the Beldens exposures in front of the escarpment of siliceous limestone. Lorraine quartz sandy shales crop out in the Yamaska River at St. Hyacinthe with dips of about 65 degrees southeastward (Foerste, 1916, p. 53-56); marine sand and clay conceals the bedrock under the intervening plain. At St. Pie, a thrust fault bounds the homoclinal sequence on the west, for the Beldens, Carman and Youngman progressively lie in contact with Glens Falls limestone. But there is no direct evidence of the magnitude of the fault—whether the succession below the Glens Falls on the west would be similar to that on the east of the thrust. The exposures of dark argillite in the river at St. Pie are somewhat west of the projection of this fault. Farther south, the dark quartz silty argillites show in several localities west of Yamaska Mountain. Probably the Highgate Springs sequence is cut out by faulting southward from St. Pie Hill and northward from outcrops about three miles north of St. Dominique, but exposure is not sufficient to permit direct observation of the relations.

AREAS OF HIGHGATE SPRINGS SEQUENCE SOUTH OF HIGHGATE SPRINGS VERMONT

Rocks like those at Highgate Springs are recognized in scattered exposures southward through Swanton and St. Albans Bay to the vicinity of Beans Point and Camp Rich, some twenty miles from Highgate Springs in northern Chittenden County (fig. 7). Invariably, Cambrian sediments lie to the east beyond the Rosenberg or Champlain fault, and dark Ordovician argillaceous and quartz silty shales are to the west. The localities have been reviewed briefly by Logan (1863, p. 855-858) and by Gordon (1923, p. 191-196, 202-204, 207-209 and 214-215); the St. Albans Bay area was crudely discussed by Edson (1906, p. 139-143) and mapped by Pretzer (1945). Little can be added by repetitious descriptions of the successive outcrops; hence their distribution is briefly listed so that the significant aspects can be discussed in subsequent pages. Most are isolated small areas exposing one formation which would appear as spots with enigmatic structure on a small scale map. The best single area to show the nature of the structure is Fonda Quarry Hill (figs. 8 and 9).

Swanton area.—The rocks are exposed in four separate areas in the vicinity of Swanton; three show only the Beldens formation. In the southeastern part of the village, Beldens white-weathering calcitite with buff-weathering dolomite interbeds is widely exposed north of the Missisquoi River; it is quarried at the Swanton Lime Works. The structure within the formation is involved in recumbent folds that can be mapped only on a very large scale and are difficult to understand in the absence of distinctive marker beds. The floor of the western quarry now filled with water, showed a nearly

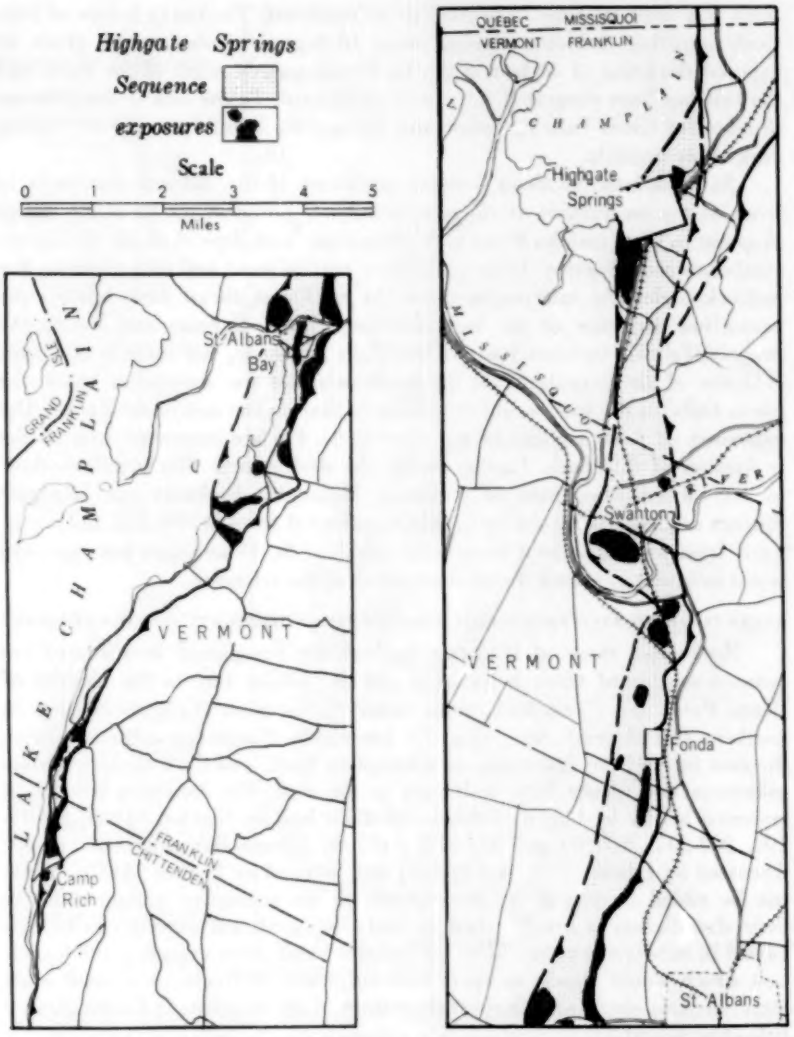

Fig. 7. Map of exposures of Highgate Springs sequence in Vermont; see figures 2 and 8 for details at Highgate Springs and Fonda.

horizontal thrust plane beneath Beldens and above black shales (Gordon, 1923, p. 192) like those extensively exposed in the river west of the village.

Half a mile east of Swanton Junction, similar Beldens carbonate rocks crop out in a small area between the St. Johnsbury and Lamoille County Railroad and Highway 104. Logan interpreted the structure as a south-pitching syncline (1863, p. 856) but this is too simple. The main exposures which

form the eastern part of the hill are principally of overturned, apparently west-dipping Beldens beds that reveal their attitude by cross-lamination and channels in dolomitized arenitic beds. The section is one of the best continuous exposures of the Beldens, but shows neither overlying nor underlying stratigraphic units.

One and one-half miles south of Swanton the Central Vermont Railroad crosses Route 104A. Just southeast of this crossing is a hill of Beldens marble apparently dipping rather constantly to the southeast. At the northwest, near the crossing, are exposures of dark slate that presumably lie below a thrust passing under the Beldens. Half a mile south of the crossing on the west side of the railroad a second hill lies between the Central Vermont and the Alburg line of the Canadian National Railroad to the southwest. This hill is dominantly of Carman quartzite, but there are limited exposures of Beldens in the cores of anticlines in the southeastern part of the hill, and the Youngman is infolded and infaulted as northeast-trending strips of argillaceous thin bedded limestones within the hill. The quartzite has been folded and broken by northeast-trending faults that are downdropped on the northwest side. In detail, the hill has quite complex patterns of outcrop of the three formations, but it is essentially of northwest-dipping Carman quartzite, broken and folded.

Fonda Junction.—The geology of the principal area of exposure west of Fonda Junction is shown in figure 8. The Lower Cambrian Dunham dolomite of the Rosenberg sequence has been thrust over a recumbent fold of Beldens limestone that structurally overlies Carman quartzite on the west; the thrust is represented by a number of planes, some enclosing lenses of Beldens within the Dunham, and the Dunham dolomite is brecciated, though not obviously so. The Carman has been thrust over Youngman, Isle la Motte and Glens Falls limestones, and the whole is torn by several faults that break the continuity of the quartzite hogback (fig. 9). This is Smiths Lime Works of Logan, who recognized the principal relations (1863, fig. 445, p. 855). Similar relations continue southward through the abandoned Rich Quarry nearer the St. Albans town line, and Glens Falls and Isle la Motte limestones crop out north of the road to the west of the Rich Quarry. On a hill about half a mile northwest of Fonda Quarry Hill are excellent exposures of upper Youngman, Isle la Motte, Glens Falls and succeeding black shale in a nearly vertical homocline on the west limb of an anticline that seems, from cleavage-bedding relations, to plunge south.

St. Albans Bay.—The escarpment of Lower Cambrian Dunham dolomite overlying the Champlain thrust stands just east of the roads leading north of St. Albans Bay to Tuller School, and south of the village to and beyond Everett School on Mill River. The thrust places the Dunham on the Beldens white limestone north of St. Albans Bay, the contact being visible within half a mile of the village. Exposures to the west and below are principally of Beldens, but the stratigraphically succeeding Carman quartzite forms quite complicated patterns including some overturned synclines—structural anticlines—along the roadside, torn by high angle faults. The Burton hills, an isolated pair of hills north of the village and east of Stevens Brook, are also

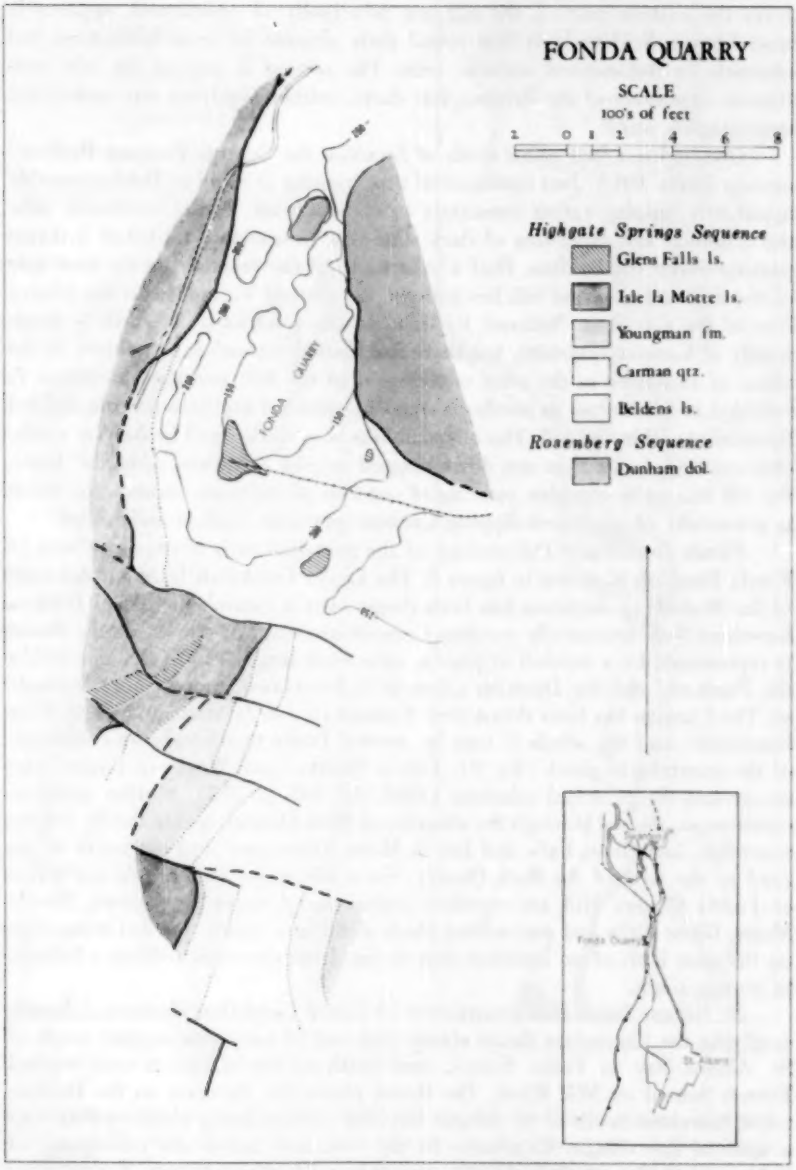

Fig. 8. Geologic map, Fonda quarry area, Vermont.

dominantly of Beldens, which has been thrust on dark slates exposed on the west slope of the western hill, the thrust plane dipping about 35 degrees to

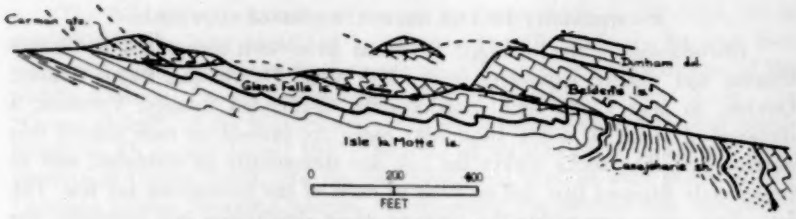

Fig. 9. Structure section through Fonda quarry.

the east; a wedge of Carman quartzite intervenes between the Beldens and the slate on the northwest slope of this hill.

The most conspicuous exposures within the village are in a north-south trending ledge of Glens Falls limestone that parallels and is just east of the shore road. Farther east near the stream are exposures of Youngman argillaceous limestone and small exposures of Carman quartzite. Still farther east is Beldens limestone, the western margin of which cuts across the trend of the younger rocks as though separated from them by a fault. The eastern exposures of the Beldens, particularly those just south of the south road, contain quartz sand and are lithically unlike other exposures seen in the slice. The Beldens continues southward along the base of the Champlain thrust escarpment through Everett School to a ridge trending southward from Lime Rock Point. Carman quartzite is exposed west of the shore road in and near Mill River, and on the shore just east of Lime Rock Point. On the point, the Beldens has been thrust over dark shales, the well-exposed surface dipping 15° N. 60° E. The thrust then trends southward toward School No. 2, for black shale is west of the Beldens ridge at the road and southward. The Dunham dolomite seems to lie directly on black shale southwestward for two miles from east of Melville Landing.

Beans Point area.—The Beldens limestone forms conspicuous exposures between the escarpment of Dunham dolomite and the black shale of the shore cliffs for two miles north and two miles south of the Franklin County—Chittenden county line, which is a mile north of Camp Rich and Beans Point. In the exposures in the slopes northeast of Camp Rich and along both sides of the road near the county line, relatively pure Beldens white-weathering gray calcilutite dips generally southeast from 15 to 60 degrees. South of the county line, the Beldens exposures form a hill above shore cliffs of slate, which lies below the gently east-dipping thrust fault. There are scattered exposures southward to Camp Watson. Throughout this narrow belt between the black shales and the Dunham, only Beldens limestone was observed.

The Champlain Fault cuts out the Highgate Springs sequence southward, so that the Dunham dolomite structurally overlies the Ordovician slates northwest of Burlington, as at Red Rock Point. The nearest exposures of similar rocks are in the northwest part of the Middlebury Synclinorium (Cady, 1945) near Weybridge, more than 30 miles to the south, and east of the Champlain thrust.

STRATIGRAPHY OF THE HIGHGATE SPRINGS SEQUENCE

Introduction.—Six lithologic divisions have been recognized in discontinuous and isolated exposures from north of St. Dominique, Bagot County, Quebec, to near Camp Rich, northwestern Chittenden County, Vermont, a distance of about 80 miles. Dissimilar rocks are present on each side of this narrow belt. The rocks within the belt are structurally so disturbed and so imperfectly exposed that full sections of each of the formations are few. The following pages summarize the sections, their similarities and contrasts, the faunas and the correlations.

Beldens limestone.—The Beldens limestone (Cady, 1945) is named from a locality in the Middlebury synclinorium north of Middlebury. The typical Beldens is dominantly fine-textured, well-bedded calcitite interbedded with buff-weathering, dark, fine-textured, brittle dolomite. The oldest rocks at Highgate Springs are Beldens, in the core of the anticline extending through Restaurant Point and the nearby islands and in the old quarry on Limekiln Point to the north. On the west side of Restaurant Point, the laminated, fine-textured, white-weathering calcitite with beds of buff-weathering dark, fine-textured dolomite is exposed in contact with the sharply contrasting quartzite of the Carman formation. In a number of small folds in the Beldens, the factured dolomite beds have been bent into regular similar folds as the adjoining calcitite has retained its continuity by flowing. The quarry in Limekiln Point exposes 50 feet of laminated calcitite, interrupted exposures of shaly carbonate rock containing quartz silt and of calcitite and dolomite extend to the exposures of Carman quartzite along the bay to the south.

The best sections of the Beldens are 1) in the lowland northwest of the escarpment west of St. Dominique, half a mile north of the highway, 2) in the hill south of Highway 104 half a mile east of Swanton, 3) west of the road to Tuller School, half a mile northeast of St. Albans Bay village and 4) on Mill River south of St. Albans Bay. The two sections near St. Albans expose the contact with the younger Carman quartzite; others are at unascertained positions within the Beldens, but they exemplify the varied rocks in the formation, and their proportions.

The dominant rock in the formation is laminated, quite pure calcitite, such as was quarried in the Fonda Quarry, where it is separated from the Carman quartzite by about ten feet of buff impure calcitite. The section north of St. Albans Bay shows that magnesium beds are not limited to the upper few feet of the 160 feet exposed below the Carman quartzite. Along Mill River to the south, 200 feet of Beldens is exposed; dolomite is virtually lacking in the upper 150 feet. In the hill east of Swanton, the formation through about 200 feet is dominantly laminated calcilutite, but contains beds of dolomitized, cross-bedded calcarenite. The section north of St. Dominique shows 120 feet of orange-weathering dark impure somewhat laminated dolomite, which tends to develop axial plane cleavage, as well as quite pure calcilutite. Similar beds are known throughout the extent of the Beldens, as in the Sudbury Nappe just north of the Taconic Mountains. The dolomite beds in the formation show excellent boudinage.

The Beldens was classed as Chazyan by Cady (1945, p. 550-552), for it succeeded rocks long identified as "Crown Point", probably because they contained poorly preserved fossils resembling *Maclurites*, characteristic of the middle Chazyan Crown Point limestone of the section in New York. The "Crown Point", of the Middlebury synclinorium was renamed the Burchards limestone (Kay and Cady, 1947) because of its structural separation from the typical Crown Point, and the Burchards and Beldens were placed in the Chipman Group, thought Chazyan. Subsequently, the Beldens was interpreted as a calcitic facies of the Bridport dolomite, the upper part of the typical Beekmantown or Canadian farther west (Kay, 1950, p. 1476), on the basis of structural continuity with the Bridport and of the presence in the Weybridge member of the Beldens of brachiopods and trilobites of Canadian affinities.

On the geologic map of west-central Vermont (Cady, 1945, pl. 10) the sequence in eastern Shoreham, Addison County, is successively Bascom-Bridport-Crown Point-Beldens-Middlebury. Along the same belt of exposure in southern Cornwall, it is Bascom-Crown Point-Beldens (including Weybridge)-Middlebury; exposures are quite discontinuous along the outcrop belt. The Bridport in east Shoreham, though dominantly of brown weathering, laminated, medium-bedded dolomite, has several beds of white-weathering calcitite; and though the Beldens formation in Cornwall is largely of white-weathering laminated calcitite, it has many beds of buff-weathering dolomite like the dominant rock in Bridport, as well as the darker gray, fine-quartz-sandy laminated, cross-laminated and thinly-bedded Weybridge member. The Burchards ("Crown Point") in Cornwall has partially dolomitized sections of fossils of the general form and dimensions of gastropods of the genus *Maclurites*, but the poorly-preserved structures may be coiled cephalopods such as *Eutrephoceras*. *Maclurites* is a long-lived genus that is abundant not only in the Chazyan Crown Point limestone, but also in the Corey limestone near the top of the Canadian in the Philipsburg Sequence near Bedford, Quebec. Rarely, specimens of *Maclurites* are present in the Beldens too, as in exposures near Camp Rich, Vermont. And *Maclurites* sp. is common in the Youngman (St. Dominique) limestone three hundred feet above the Beldens on St. Pie Hill and northward in Bagot County, Quebec.

Not only does the Beldens seem to pass laterally into the Bridport from Cornwall to Shoreham, but the Weybridge member of the Beldens contains fossils that seem diagnostic of Canadian age. These fossils have been collected most successfully in the quartz-sandy laminated dark calcitites of the Weybridge east of the road east of "The Ledges" on the Cornwall-Weybridge town line two miles west of Middlebury. The fossils are generally poorly preserved, for the quartz grains have grown and penetrated the calcite. Species of the brachiopod genera *Diparelasma*, *Orthambonites* and *Syntrophopsis*, and of the trilobites *Goniurus* and *Isoteloides* have been identified, forms that are widely distributed in Upper Canadian sediments (Ulrich and Cooper, 1938).

In the Highgate Springs sequence, the Beldens is the oldest stratigraphic unit that has been identified confidently. In a few localities there are thin-

bedded quartz silty limestones resembling the Weybridge facies—above the quarry on Limekiln Point near Highgate Springs, and in the northwestern exposures near Highway 104 east of Swanton. But recognizable fossils have not been seen. A small area of calcitite containing coarse quartz sand associated with the Beldens, either within or below the formation, lies south of the road just east of St. Albans Bay Village. The sandy beds have erosional disconformable contacts and contain orthoceracone cephalopods of undetermined identity.

The exposures of equivalent rocks that lie nearest to the Highgate Springs sequence are on "The Head" at the south end of Isle la Motte, ten miles directly west of St. Albans Bay. The Bridport, there underlying the basal Chazyan Day Point sandstone, consists of tan-weathering, very dark gray, laminated, medium-bedded somewhat siliceous dolomite, with very thin-bedded argillite layers. This sort of rock is again present on Providence Island off southern South Hero, Grand Isle County, Vermont, and in western belts in west-central Vermont (Cady, 1945). As stated above, the Bridport dolomites pass laterally eastward into the Beldens calcitites with interbedded dolomites. In the Highgate Springs sequence, the only section in which the Bridport facies dominates through a considerable thickness is that northwest of St. Dominique. The nearest exposures to the west are on Ile Jesus and near Chateauguay in the Montreal area (Clark, 1952), about 40 miles west of St. Pie. The Canadian Beauharnois formation is dominantly of bedded dolomite and argillite in the thickest sections, but the youngest fossil-dated beds are thought to be older than the Bridport. The Canadian beds are known also from several wells between the Montreal outcrop and the St. Dominique belt (Clark, 1955, p. 8).

The thickness of the Beldens is determined only as a minimum, the two hundred and fifty feet of so measured in continuous sections; the thickness may be considerably greater and is estimated as 600 to 700 feet on the east limb of the Middlebury Synclinorium to the southeast (Cady, 1945, p. 551); the Bridport dolomite has an estimated thickness of 470 feet in East Shoreham, west of the synclinorium (Brainerd and Seely, 1890, p. 3).

The most striking fact about the Beldens in the Highgate Springs slice is that it is persistent throughout the belt of exposure, yet in no place has the underlying Burchards formation been definitely recognized; even if the quartz-sandy calcitite near St. Albans Bay be older than Beldens, the area of exposure is very small. Yet the Beldens commonly lies directly beneath the Lower Cambrian rocks of the Rosenberg slice, as well displayed in Fonda Quarry, and as essentially true southward to Camp Rich. The Beldens has been severely folded and has flowed; not only are small folds frequently recognized by the course of dolomite beds, but the calcitite thins across the swells in dolomite boudinage structures. The formation was relatively incompetent and served as a glide-plane along which the thrust advanced.

Paleogeographically, the Beldens is a facies of the Bridport. On a paleolithologic map dolomites would dominate in a belt extending westward from west of the west limb of the Middlebury Synclinorium, cutting [illegible] outcrop about in eastern Shoreham.

Carman Quartzite.—The Carman quartzite has its type section at the head of Shipyard Bay in the Highgate Springs area (Kay, 1945); it contains about 120 feet of thin- to medium-bedded quartz sandstone having a few argillaceous partings, and includes some calcarenite beds in the upper part. The section exposes the contact with the underlying Beldens calcitite, but the top is cut off by a fault. The quartz sandstones are somewhat argillaceous; surfaces commonly show rod-like structures, "fucoids". In other sections, the upper part of the formation is varyingly argillaceous. The type section is the thickest that has been seen, even though incomplete. The formation is completely exposed in Stanbridge Township, where it is but 40 feet of quartzite passing up into argillites with nodular calcareous lenses. On western St. Pie Hill, it is about 120 feet thick; the lower half is quartzite, the upper half, quartz silty argillite with quartz sand and calcarenite beds. Only the uppermost beds are exposed west of St. Dominique, but calcarenites have yielded a small faunule. Southward from Highgate Springs, good exposures at Fonda Quarry show about 40 feet of quartzite, and many scattered bands are associated with the Beldens north of St. Albans Bay and southward to the cove east of Lime Rock Point.

The Carman quartzite is like the quartz sandstones and associated argillites in the base of the Day Point formation of Grand Isle County, Vermont, to the west (Oxley, 1952). Moreover, the faunules from the immediately succeeding limestone at St. Dominique have many specimens of *Raphistoma striata* (Emmons), abundant in lower Chazyan Day Point calcarenites in New York and the islands of Lake Champlain (Oxley, 1952; Raymond, 1906).

In the northern Middlebury Synclinorium, quartzite is not constantly present in the basal part of the Middlebury formation, but quartz-sandy beds are common, as in exposures north of Center Weybridge, three miles northwest of Middlebury. In the Sudbury Nappe at the northern end of the Taconic Range, either the Middlebury argillicalciutite lies directly on Beldens calcitite, or else they are separated by as much as 50 feet of partly dolomitized, thin-bedded to medium-bedded, somewhat argillaceous calcisiltite, which forms conspicuous hills around Sudbury and in which axial-plane cleavage is well developed as magnificently shown on "Recumbent Hill", west of the Lemon Fair River a mile west of Sudbury church.

Youngman formation.—In the type section at Highgate Springs, Vermont, the Youngman formation (Kay, 1945) has exposure of a lower 180 feet of argillaceous calcitite and argillite lying on Carman quartzite, and beyond interuption an upper 30 feet of argillite with calcareous beds below Isle la Motte limestone (fig. 4). The full thickness in this locality is probably more than 300 feet, and there are several quite fossiliferous horizons.

In some respects, the type section is unique. Other sequences in the same apparent stratigraphic position along the thrust slice are not lithically identical, though varyingly similar, and in Vermont they are virtually non-fossiliferous. East of St. Hyacinthe, Quebec, the St. Dominique limestone (Clark, 1955) in the same stratigraphic relations is more calcareous and quartz-silty and has common fossils. The type Youngman contains some fossils known in the Chazy limestones to the west along Lake Champlain, but

others that are unknown there and have been thought diagnostic of later age; hence the classification of the Youngman as Chazyan has more than local interest.

In the type section of the Youngman along the lake shore north of the village, the lowest 150 feet is dominantly argillaceous calcisiltite and calcilutite with partings and thin beds of argillite. The succeeding exposures are dominantly argillite, but with thin beds of calcitite. The rocks have been deformed so that argillaceous matter has penetrated calcareous beds along cleavage, making them cobbly, or producing slaty cleavage that is more evident than the bedding; thus the field geologist tends to call the formation a slate.

The only other continuous exposure of considerable thickness is in a hill .7 mile north and a little west of Fonda Quarry, south of Swanton, where 80 feet of section extends to the top of the formation. These beds bear little resemblance to the upper Youngman at Highgate Springs, for the section is dominantly of well-bedded calcisiltite, though with argillaceous partings and several feet of argillite with calcareous cobbles in the lower 20 feet. There are many scattered exposures of smaller parts of the formation, most of them of cleaved calcareous argillite as in the two good sections. To the north in Quebec, the basal few feet, argillite with calcareous interbeds, crop out in Stanbridge. The sections in the same stratigraphic position at St. Pie Hill and St. Dominique are discussed below as the St. Dominique limestone.

The thickness at Highgate Springs can be computed as 285 feet, but is greater, for a fault crossing the section cuts out beds along the strike; the true thickness is probably greater than 300 feet.

The Youngman formation is ordinarily non-fossiliferous, but fossils have been found in thin lenses and beds at several horizons at Highgate Springs. *Lonchodomas halli* (Billings) is abundant in faunules from both the basal 20 feet and the upper 50 feet. *Christiania* sp. cf. C. *subquadrata* (Hall) is abundant in the lowest beds. The associated forms are listed in a table; G. A. Cooper has been helpful in the identification of the brachiopods.

St. Dominique limestone.—The St. Dominique formation was described (Clark, 1955, p. 9) as consisting of the St. Dominique sandstone member, about 110 feet thick at St. Pie Hill, and the St. Dominique limestone member, 225 feet in the same section. The writer has earlier (1945) considered these to be the Carman and Youngman formations, but the differences are sufficient to make separate designation preferable. The Beldens is succeeded by 120 feet of argillite with quartz arenite dominant in the lower part and calcarenite beds frequent in the upper; These are overlain by about 200 feet of carbonate rocks, principally calcarenite and calcisiltite, partially dolomitized, to a concealed interval. The writer computed a thickness of 340 feet from the Beldens to the "Black River," almost identical to the 330 feet given by Clark.

The other section of the St. Dominique is along the highway northwest of St. Dominique. The basal sandstone is incompletely exposed but the succeeding limestones are extensively quarried through a thickness of nearly 300 feet, principally of somewhat siliceous calcisiltite, but with calcarenite prevalent particularly in the upper part.

TABLE 1

Faunules from Youngman Formation, Highgate Springs, Vermont

Localities:	a	b	c	d
Christiania subquadrata (Hall)	a*	.	.	.
Hesperorthis ? sp.	f	.	u	.
Leptellina sp.	c	.	.	.
Lingulella ? sp.	u	.	.	.
Platymena sp.	f	u	.	.
Protozyga sp.	.	r	.	.
Rostricellula sp.	.	.	u	f
Skenidioides ? sp.	.	.	r	.
Sowerbyella sp.	.	.	c	.
Valcourea sp. cf. *V. strophomenoides* (Raymond)	.	.	c	.
Maclurites sp.	.	r	.	.
Orthoceracone	x	.	.	.
Bathyurus sp.	.	.	f	.
Ceraurus ? sp.	r	.	r	.
Calliops sp.	.	.	u	.
Illaenus sp.	u	.	.	.
Isotelus sp.	.	.	c	.
Lonchodomas halli (Billings)	c	.	a	.
Onchometopus sp.	.	.	c	.
Remopleurides sp. cf. *R. americanus* Raymond	r	.	.	.
Lepidocoleus sp.	r	.	.	.
Aparchites sp.	c	.	c	.
Eridoconcha sp.	.	.	r	.
Euprimitia sp.	f	.	.	.
Eurychilina sp.	.	.	r	.
Schmidtella sp.	c	.	.	.
Thomasatia sp.	.	.	r	.
Winchellatia sp.	.	.	r	.

Localities:—along shore of Missisquoi Park from Kingfisher Bay to Phelps Bay; a- basal 20 feet; b- succeeding 150 feet, from several beds; c- single 1-inch lens, 27 feet from top of section; d- bed 15 feet from top of formation.

* Frequencies: a- abundant, c- common, f- frequent, u- uncommon, r- rare, x- present.

Fossils are frequent in the lowest calcarenites exposed above the quartzite south of the highway, and in calcarenites near the top of the formation north of the large quarry north of the road (table 2). At St. Pie Hill, the upper St. Dominique limestone has reefy beds in which *Eospongia* and *Maclurites* are frequent, much as in the upper beds of the middle Chazyan Crown Point formation in New York. The lowest beds at St. Dominique have abundant *Raphistoma stamineum* Hall; though this form is rather distinctive of the lowest Chazyan Day point formation in New York, virtually identical forms have long range.

Classification of the St. Dominique and Youngman formations.—The exposures at St. Dominique have invariably been classified as Chazyan by those who have studied them since the time of Logan (1863, p. 206), and have been correlated with the exposures at Highgate Springs (p. 820; also Raymond, 1906, p. 555). Though *Lonchodomas halli* was originally described from Highgate Springs and has been reported from St. Dominique and the typical Chazy of New York (Raymond, 1905, p. 334) through considerable range, *Christiania subquadrata*, common in the lower Youngman at Highgate

TABLE 2

Faunules from St. Dominique limestone, Bagot County, Quebec

Localities:	1a	1b	2a	2b
Bryozoans	x*	x	.	x
Eospongia sp. cf. *E. roemeri* (Billings)	.	.	.	f
Camerella sp.	.	r	.	.
Hesperorthis ? sp.	u	.	.	.
Platymena sp.	c	.	.	.
Sphenotreta sp.	.	c	.	.
Valcourea sp.	c	a	c	.
Bucania sulcatina (Emmons)	.	.	r	.
Raphistoma stamineum Hall	c	r	c	.
Maclurites magnus Leseuer	.	.	.	x
Bronteopsis ? sp.	.	r	.	.
Bumastus sp.	.	.	x	.
Calliops sp.	.	f	.	.
Ceraurinella sp. cf. *C. pompilius* (Billings)	.	.	r	.
Illaenus sp.	.	r	.	.
Isotelus sp.	c	.	c	.
Platymetopus sp. cf. *P. minganensis* (Billings)	.	u	.	.
Pseudosphaerexochus sp.	r	.	.	.
Remopleurides sp.	r	.	.	.
Aparchites sp.	c	c	.	.
Bythocypris sp.	.	r	.	.
Dilobella ? sp.	r	.	.	.
Eridoconcha sp.	.	r	.	.
Euprimitia sp.	f	.	.	.
Eurychilina sp.	f	f	.	.
Schmidtella sp.	.	f	.	.
Cystid fragments	a	a	.	c

Localities: St. Dominique: 1a-calcarenite immediately above quartzite south of highway at west face of quarry hill; 1b- calcarenite ledges about 50 feet from top of formation northeast of quarry north of highway. St. Pie Hill: 2a- ledges in upper part of formation near north end of hill; 2b- bioherms on northeast slope of hill.

* Frequencies: a- abundant, c- common, f- frequent, r- rare, x- present.

Springs, seems to have been discovered first by the writer. Experience with this genus and its associates had led to the judgement that in the Appalachian region *Christiania* is post-Chazyan, and it is so indicated on the Ordovician Correlation Table (Twenhofel and others, 1954) though Rodgers (*ibid.*, p. 267) suggests that—"*Christiania* ranges even lower than shown on the chart into equivalents of the Chazy". Those who have studied brachiopods most diligently have been reluctant to accept the Chazyan age of the Youngman formation (p. 287). This becomes further involved in the relative age of the "upper Lenoir" *Christiania* zone to the Lincolnshire limestone, which Rodgers places above the zone and Cooper below it. B. N. Cooper (*ibid.*, p. 274) tentatively correlates the Lincolnshire with the upper Chazyan, noting that "*Sowerbyella, Sowerbyites,* and *Dinorthis* are unknown in the type Chazyan of the Champlain Valley"; the writer has classified the Lincolnshire as upper Chazyan (1956, p. 98), because of the similarity of the *Billingsaria* reefs in the Lincolnshire and Valcour, as well as approximate stratigraphic position.

The stratigraphic relations of the Carman and Youngman seem to preclude their being other than Chazyan. If they were two similar lithic sequences, each a sandstone-limestone couple, they would have to persist in contiguous areas, yet each be absent where the other is present. This contrast would have to take place not only from the Highgate Springs sequence to the foreland, but also between the St. Dominique and the Carman-Youngman along the same trend; the Middlebury would continue this enigma.

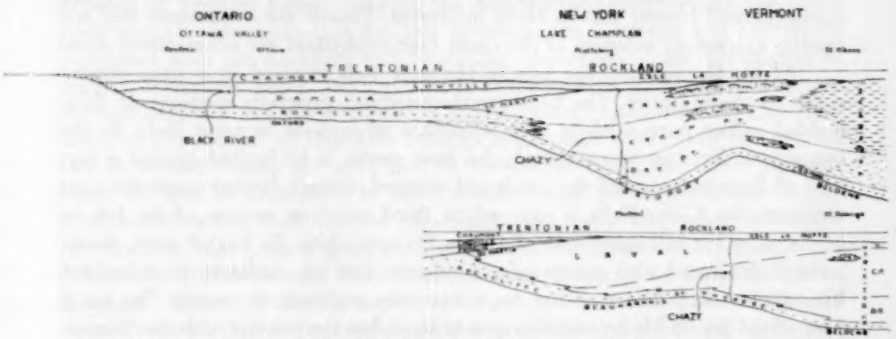

Fig. 10. Restored sections of the Chazyan rocks from the Ottawa Valley to Quebec and Vermont.

The conclusion reached by the writer is that the Carman-Youngman is Chazyan of a facies different from the typical Chazy group (fig. 10) and from the St. Dominique, and that the presence of *Christiania, Sowerbyella, Platymena* and other forms in the formation must be accepted as showing that the ranges of these genera are somewhat greater than had been recognized, and that their sensitivity in correlation is somewhat reduced. It is unfortunate that the fossils are so rarely present, and not preserved so that they might be etched and better examined. The writer has tried material by several methods of preparation, and found none that is better than tooling them from the calcite matrix.

Isle la Motte limestone.—The Isle la Motte limestone was named by Emmons (1842) from the island town of Grand Isle County, Vermont, in northwestern Lake Champlain; he correlated it with the "Seven-foot Tier", the upper part of the Chaumont limestone of the Black River group of northwestern New York. Though the writer considered it lower Trentonian (1937, p. 260), he is inclined to consider this in error, though the evidence is somewhat contradictory; the question is discussed below. In the Highgate Springs sequence, a similar limestone persists between the Youngman formation and the Glens Falls limestone; the "Lowville" calcilutite that separates the typical Isle la Motte from the Chazyan is not present.

The formation is everywhere thick-ledged, though medium-bedded, resistant calcitite and tends to form ridges, particularly where the Youngman is argillitic as at Highgate Springs; black chert is locally present. The full thickness of the formation is shown on the south side of Phelps Bay, Highgate Springs, where about 70 feet is exposed; the lowest beds contain quartz

sand and lie with nearly a foot of erosional relief on argillaceous Youngman containing nodular calcilutite beds. The base is similar where exposed on the southwest side of the U-shaped loop of its outcrop west of the village. The contact with the succeeding Glens Falls is excellently exposed on the west side of the hill directly south of the village crossroad, the lithology changes abruptly into the more argillaceous, thin-bedded fossiliferous Glens Falls. The Isle la Motte is generally calcisiltite, with some well-indurated calcarenite interbeds and lenses; rarely, chert is present. Fossils are uncommon and not readily extracted; colonies of the coral *Foerstephyllum* are occasionally seen.

At St. Dominique, the Isle la Motte is well exposed in a lime quarry south of the highway. The beds are thick-ledged, generally medium- to thin-bedded rather pure calcitite with abundant pelecypods in some beds. In the quarry, 48 feet can be measured; the base seems to be faulted against a surface of Youngmans, and the top is sod covered, though farther south the contact with the Glens Falls is exposed. A third excellent section of the Isle la Motte is in the hill northwest of Fonda Quarry where 70 feet of dark, thick-bedded, fractured and calcite-veined calcitite lies on similarly thick-bedded Youngman calcisiltite that has more common argillaceous seams. The basal part of the Isle la Motte contains quartz sand, but the contact with the Youngman is not striking. Moreover, the upper contact with the medium- to thin-bedded Glens Falls is not conspicuous, though the latter is distinctly more argillaceous and is fossiliferous rather than essentially barren.

With regard to correlation and classification, this unit is lithically like the Isle la Motte limestone exposed in belts to the west. The typical Isle la Motte overlies light-weathering calcilutite that is presumably "Lowville", which in turn lies disconformably on the Upper Chazyan Valcour limestone. The Isle la Motte is succeeded by thin-bedded argillaceous Glens Falls limestone, commonly fossiliferous, and the fossils seem Kirkfieldian in the type area. Thus the typical Isle la Motte could be Blackriveran Chaumontian or lowest Trentonian Rocklandian on the basis of stratigraphic position, if the correlations of the Lowville and Glens Falls are correct. The fossil considered most diagnostic of upper Rocklandian is *Triplesia cuspidata* (Hall), found below the Glens Falls at Crown Point (Raymond, 1903, p. 23; Kay, 1937, p. 261); the writer considered these beds to be in the Isle la Motte. However the sixteen feet of limestone below the Glens Falls at Crown Point is calcarenite and calcisiltite that is softer, less brittle and coarser textured than the underlying beds; as the latter are like typical Isle la Motte, this opposes classification of the Isle la Motte as Rocklandian. The Isle la Motte contains species of *Receptaculites*, *Maclurites* and *Foerstephyllum* like those particularly common in the Rockland in Ontario; but similar or indistinguishable forms may well be present in the Black River equivalents, particularly as the lithologies, and presumably the ecologies represented, are so similar that rocks of these ages are frequently confused. The writer was further influenced by finding "*Rafinesquina*" sp. and "*Calymene*" sp. in the "Isle la Motte" of Highgate Springs, which lies directly on Chazyan-classified Youngman rather than on "Lowville"; these genera are not known below Trentonian in western New

York; perhaps they ranged lower than has been supposed, and moreover I suspect misidentification.

In summary, the Isle la Motte formation is now classified as Blackriveran, and correlated with the Chaumont limestone of the Black River group of northwestern New York.

Glens Falls limestone.—The Glens Falls limestone was named by Ruedemann (1925, p. 22) for the "basal Trenton limestone" . . . "lower than any exposed at Trenton Falls", between the Amsterdam limestone and the Canajoharie shale. The formation has two members, the Larrabee and Shoreham, correlated with the Kirkfield and Shoreham limestones of the Trentonian of western New York (Kay, 1937, p. 262-267; Chenoweth, 1952). The Highgate Springs sequence has many exposures of beds faunally and lithically like those south and southwest of Highgate Springs crossroads, where they succeed the Isle la Motte; fossils are abundant in several horizons, and show that the lower 35 feet is equivalent to the Larrabee, and the higher *Cryptolithus-Reuschella*-bearing beds, to the Shoreham. The beds are rather uniformly thin-bedded, black, dark gray-weathering, quite argillaceous calcilutite and calcisiltite, with occasional beds of coquinal calcarenite. The thickness can be measured in the hill directly south of the spring east of the railroad, for the discontinuous vertical-dipping exposures are flanked by Isle la Motte and Hortonville; it is almost 100 feet. The higher part of the Glens Falls and the succeeding black shale are exposed along the southwest shore of Shipyard Bay. There are east-dipping exposures of the Glens Falls in the fields east of the Isle la Motte limestone quarries at St. Dominique. The formation is quite fully exposed with nearly vertical dip on the west slope of the hill northwest of Fonda Quarry, and there are scattered exposures along the road west of Rich quarry, as well as in St. Albans Bay village.

The abundant fauna of the Glens Falls permits it to be correlated readily with the formations of western New York. *Reuschella edsoni* (Bassler) was first described from the Shoreham beds at Highgate Springs; it has been seen elsewhere only in the section at Larrabee Point in Addison County, and in exposures in hills one mile west of Hyde Manor, in calcarenites directly beneath the typical Hortonville. The genus was described originally from the Caradocian on the Onny River, Shropshire, England, and has been found to be common in the Oranda formation in the Appalachians of Pennsylvania (Craig, 1949) and the Virginias (Cooper and Cooper, 1946). *Cryptolithus tesselatus* Green, on the other hand, characterizes a single limestone unit called the Shoreham as far westward as southeastern Ontario (Chenoweth, 1952, p. 526.)

The Glens Falls limestone seems rather constant throughout the area east of Lake Champlain, whether in the Highgate Springs slice and the Middlebury Synclinorium, or in the unthrust terrane to the west. The formation is directly succeeded, however, by black shaly argillutite in the overthrust sequences, but by the interbedded argillaceous calcisiltites and argillutites of the Cumberland Head formation along Lake Champlain (Hawley, 1957). Inasmuch as the name Hortonville was applied to shales directly succeeding Glens Falls, the term is applied in the Highgate Springs sequence. The Stony

Point black shales that overlie the Cumberland Head formation along Lake Champlain to the west are presumably somewhat younger at their base.

Hortonville Shale.—The Hortonville shale (Keith, 1932, p. 360; Cady, 1945, p. 558) overlies the Glens Falls limestone. The laminated black argillaceous shale is exposed only in limited areas within the slice, and no section is continuous for more than a few feet. The basal few feet tend to contain thin calcilutite beds, but they disappear upward into dark shale. Only rock that is in rather continuous exposure and structure with older formations in the sequence can be confidently attributed to the Hortonville in the slice. For example, at Lime Rock Point and southward, to the southwest of St. Albans Bay, the Beldens white-weathering calcitite is thrust on dark shales, but the latter are in the Iberville formation, and may be as much as half a mile stratigraphically above the base of the Hortonville shale (Hawley, 1957). The recognition of the great stratigraphic difference between the shales in the Highgate Springs sequence and those that lie across the Highgate Springs fault on the west emphasizes the significance of the fault. The youngest preserved shale formation to the west, the Hathaway, contains radiolarian cherts on Hathaway Point in St. Albans Bay, but also blocks of Glens Falls limestone which lies half a mile below and must have been elevated in a structure to the east in Hathaway time.

On the east, there are no problems where rocks bound the Highgate Springs sequence from its southernmost exposures in Chittenden County to the southern and undetermined limit of the Philipsburg sequence in the vicinity of Swanton. From here to the latitude of Bedford, the Philipsburg sequence contains formations that can be confused only with the Beldens formation in the Highgate Springs sequence, if with that. Northward, the carbonates of the Philipsburg sequence are succeeded in the axis of a northward pitching synclinorium by the Mystic limestone boulder conglomerate and the dark, laminated argillites of the Stanbridge shale, which are middle Ordovician and much resemble the shales of the Highgate Springs sequence. The Philipsburg thrust can be traced readily northward to the latitude of Bedford, where the fault, trending more northeasterly than the axis of the synclinorium, brings Stanbridge slate over Iberville and Stony Point shale of the more westerly sequence, the Highgate Springs Sequence being cut out. The structural position of the slates exposed at Farnham and St. Pie has not been determined. Just to the north, on St. Pie Hill, the Beldens has been thrust over the Glens Falls limestone, but this may be a fault within the Highgate Springs sequence. Cincinnatian Lorraine sandstone and shale is well exposed in the river at St. Hyacinthe, but there is no exposure in the several miles southeastward to the Beldens northwest of St. Dominique. Thus, shales known to be Hortonville because of structural continuity with Glens Falls in the Highgate Springs sequence can adjoin similar Stony Point, Iberville or Hathaway shales west of the Highgate Springs fault, and similar Stanbridge shales east of the Philipsburg fault.

STRUCTURAL GEOLOGY

The structural relations of the sequence have been briefly discussed. The Rosenberg sequence lies next east in the hanging wall of the Champlain or Rosenberg thrust, from Beans Point, Chittenden County, Vermont to the vicinity of Swanton. The fault contact is beautifully displayed in Fonda Quarry and can be seen or closely approached for miles to the southward. The strata that adjoin the Highgate Springs sequence are invariably Lower Cambrian, and generally within the Dunham dolomite, a relatively competent, gently east-dipping thick-bedded reddish dolomite that succeeds Cheshire quartzite in normal sequence on the east limb of the Middlebury Synclinorium. Thus the Rosenberg thrust has clearly moved at or above the contact between the Cheshire and the Dunham.

The Rock River dolomite of the Philipsburg sequence overlies the Philipsburg thrust where it bounds the Highgate Springs sequence north and south of the Vermont-Quebec boundary, or, as at Philipsburg, cuts over onto the sequence normally west of the Highgate Springs (McGerrigle, 1930). Northward in Quebec, the trace of the Philipsburg Thrust cuts upward in the stratigraphic section; the Corey limestone, which lies a quarter mile east of the Highgate Springs sequence exposures in Stanbridge Township, is nearly 2000 feet higher in section than the Rock River. The formation on which the Rock River lies is not exposed, but by analogy to similar sequences in Vermont, it should be heavy-ledged carbonate rock. Northward, it is presumed that the Philipsburg thrust continues unrecognized between Stanbridge shale on the east and Hortonville shale on the east flank of the St. Pie-St. Dominique homocline.

The most striking features of the Highgate Springs sequence are the persistence of a group of formations of mixed types and only a few hundred or a thousand feet thickness for such a long distance in such a narrow strip, and the constant presence of the Beldens high-calcium limestone interbedded with dolomite as the oldest rocks in the thrust slice. The underlying formation must originally have been the Burchards, somewhat dolomitized calcitite, or the Bascom, argillaceous limestone, partly rather shaly, but commonly medium-bedded. Whatever the cause, the Highgate Springs slice seems to have moved on a thrust surface within or about the base of the Beldens formation. The effect of the proximity of the thrusts above and below is evident in the severely folded and fractured sediments, as at Highgate Springs and Fonda Quarry.

The several formations have reacted differently. The Beldens calcitite flowed readily, the interbedded dolomite being drawn out into lens-sectioned slivers or boudins. The Carman has been rather severely folded in spite of being quartzite but it commonly also has gash or tension fractures normal to the surfaces of the beds. The Youngman, where shaly and argillaceous, has developed axial-plane cleavage to such a degree that the cleavage surfaces, along which the argillite has flowed, come to be quite as obvious a series of parallel bands as the bedding; the two can easily be confused. The Isle la Motte is more resistant and tends to be crushed and to develop calcite-filled

veins. The Glens Falls develops excellent axial-plane cleavage, which causes the rock to break into rhomb-shaped cleavage fragments. The Hortonville, when well cleaved, is a slate.

The Highgate Springs slice is essentially a torn and deformed segment from the overturned east limb of a northeast-plunging recumbent synclinorium cored by Ordovician chert and shale in Vermont (Hawley, 1957). The sole of the Highgate Springs slice, the Highgate Springs thrust, has a more northerly trend than the northeast-trending axis of the synclinorium. In Quebec, the trace of the Highgate Springs Thrust cuts the east limb of the southern end of the Chambly-Fortierville synclinal basin (Clark, 1947).

The principal structure to the east in Vermont is the northward-plunging synclinorium of Cambrian sediments having its west limb in the Rosenberg slice and its east limb in the Oak Hill sequence (Booth, 1949). At the international boundary and northward, the Rosenberg slice is separated from the Highgate Springs slice by another north-plunging syncline, the Philipsburg slice. The Rosenberg thrust is a bedding plane slip on the lower Cambrian, presumably on some relatively incompetent horizon beneath the Dunham dolomite, which crops out constantly at the base for scores of miles in Vermont and Quebec. The displacement of the Rosenberg slice over the Philipsburg was sufficient for the Rosenberg slice to cut over progressively younger beds northward, but the Philipsburg sequence being almost wholly Ordovician, and the Rosenberg Cambrian, there is little to compare stratigraphically between them. On the other hand, the Philipsburg sequence has strata of the same stratigraphic level as but lithologically distinct from, the Highgate Springs sequence and must be rather far travelled, as the Highgate Springs sequence is comparatively constant for several scores of miles. The strongest stratigraphic similarities of the latter are with sediments in the northern Middlebury synclinorium, a south-plunging structure in the southern continuation of the Rosenberg slice block, above the Champlain-Rosenberg thrust.

Recent studies in the synclinorium west of the Highgate Springs slice suggest that folding began in Trentonian time. Submarine breccias in the Hathaway formation in the core of the synclinorium (Hawley, 1957) contain fragments of Glens Falls limestone, which lies far below the Hathaway in normal sequence. Hence, a welt of Ordovician must have risen to the east of the synclinorium and yielded the Glens Falls boulders. As the Glens Falls is present in the Highgate Springs slice and in the Middlebury synclinorium, but absent in the Philipsburg Sequence, the source may have been in an arch rising on the east flank of the Highgate Springs belt.

CLASSIFICATION OF THE ORDOVICIAN SYSTEM

Introduction.—The stratigraphy and fauna of the Highgate Springs sequence bears on current classifications of the post-Canadian—pre-Trentonian Ordovician. To avoid ambiguity in nomenclature, it is considered that the Canadian Series extends to the top of the Bridport and Beldens formations in the Champlain Valley, and that the Trentonian Series has its base at the contact of the Selby member of the Rockland formation on the Chaumont

limestone of the Black River Group in northwestern New York and eastern Ontario.

Principles in terminology.—Stages are time-stratigraphic units defined by rocks in a place. Lithically distinct units are formations or members. Some units called formations were defined as stages. Thus Raymond's (1914, p. 348) "formations" in the Trenton were zones he thought time-defined by fossils; subsequently the lithologies were distinguished, but Rockland limestone and Rocklandian stage are identical at Rockland, Ontario. Other formations in the stage have lithic differences or were named as an incident to their isolation. Cushing's divisions of the type Chazy were described as substages (1905, p. 368).

It is better to give a new name than to redefine an old one that has clearly stated and established content, unless the proposed changes are minor or a reflection of ambiguity. The Chazyan Series corresponds to the Chazy limestone as originally used for rocks above the Calciferous and below the Black River near Lake Champlain. The Trentonian extends from the top of the Black River in northwestern New York to the top of the Utica. The Bolarian Series includes only rocks younger than the Lincolnshire and older than the Nealmont in Virginia (Kay, 1947). If such terms are inapplicable to spans that should have names, they should not be redefined. There may be a little inconvenience in the introduction of new terms, but that is better than confusion.

Ordovician Classification.—Classification becomes involved in judgements of correlations unless the full sequence is in one locality. The classification of rocks younger than Canadian and older than Trentonian concerns sequences a) along Lake Champlain, containing the type Chazy, b) in northwestern New York, containing the typical Black River and Trenton, and c) in the Appalachians, particularly the Virginias, where there are type sections of the Bolarian Series and of a number of recently named stages (Cooper, 1956, p. 8). Classification and correlation discussed within the past two years (Cooper, 1956; Kay, 1956; Twenhofel and others, 1954) must be reconsidered.

Champlain Valley and northwestern New York.—The Chazy limestones lie on the Canadian along Lake Champlain; now that the Canadian age of the Beldens limestone has been recognized, there is no problem in the placement of the boundaries. The Chazy lies on a regional unconformity with an overlapping basal quartz sandstone. The Chazy has long been divided into three lithologic and faunal zones (Brainerd and Seely, 1888) which were given the "substage" names Day Point, Crown Point and Valcour (Cushing, 1905; as stages these have been called Dayan, Crownian and Valcourian (Oxley, 1952).

The Black River group of northwestern New York lies on Precambrian crystalline rocks or Cambrian (Young, 1943). The original Black River has come to be divided into the Pamelia, Lowville and Chaumont formations in northwestern New York, but younger beds have been called "Black River" in Ontario (Kay, 1942, p. 596; 1956, footnote, p. 96). The limestones that succeed the Chazyan along Lake Champlain are assumed to be correctly cor-

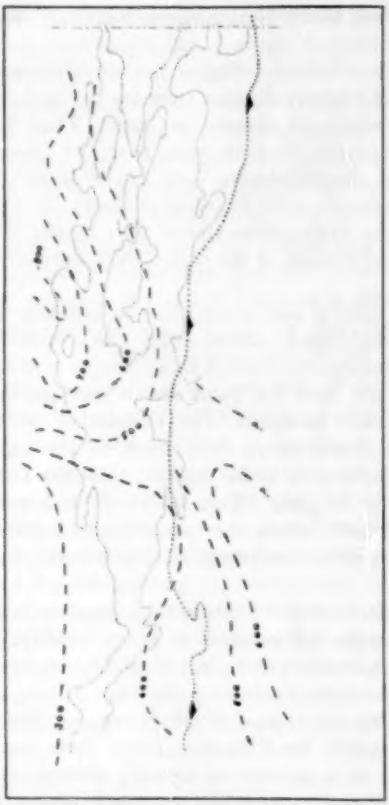

Fig. 11. Thickness of Chazyan rocks along Lake Champlain in New York and Vermont, based largely on data from Oxley, 1952.

related with the Lowville and the Chaumont (Isle la Motte); the Lowville is interruptedly present. Thus the upper Black River is younger than the type Chazyan. In the Ottawa-St. Lawrence Lowland (Wilson, 1946), the St. Martin limestone has been classified repeatedly as Chazyan Valcourian; it underlies Pamelia beds. Thus if the generally accepted correlations are correct, the Chazyan Series is older than some or all of the Pamelia formation of the Black River group of northwestern New York.

The Appalachian Region.—The Appalachian region has been thought to have fuller sequences of post-Canadian rocks than New York, but the recent studies cast some doubt on this assumption. If one established an Appalachian standard, one could correlate to New York sequences; there are problems of correlation within the Appalachian sections, so a single one must be used. The top of the Canadian seems marked by a readily recognized disconformity in most places. The writer considers the top of the Black River equivalents to be the Stover limestone in Pennsylvania and the McGraw limestone in the

Virginias (Kay, 1956, p. 97). To the southwest, there is disagreement whether the base of the Trentonian is within the upper Witten formation or at the top. Cooper (1956) placed the lowest Trentonian as well as Black River equivalents in his Wilderness stage.

The Bolarian Series was defined (Kay, 1947) as extending from the top of the Lincolnshire limestone to the base of the Nealmont limestone, which seems lowest Trentonian. The Lincolnshire was classified as Chazyan.

Rocks younger than Canadian in the Appalachian region have been assigned by Cooper to the Marmor, Ashby, Porterfield and Wilderness stages (1956, p. 8). The Marmor, based on formations in Tennessee "includes the Chazy group of rocks and its correlatives." As the type section of the Marmor is isolated from that of the others, I will refer to the rocks between the Canadian and the Elway and Lincolnshire formations of the type Ashby in Virginia as the Blackford "stage." Thus the Virginia sequence consists of the Blackford, Ashby, Porterfield, (containing Ward Cove, Peery and Benbolt formations) and Wilderness "stages"; the writer prefers that stage names have "an" or "ian" endings, but will apply quotation marks to distinguish "stages" from formations. There is question of their correlation with the "Marmor." The essential problem is that of the time correlation of these with the Chazy and Black River in New York.

Bearing of the Highgate Springs Sequence.—The stratigraphy and faunas of the Youngman formation and St. Dominique limestone bear directly on the classification. If they are Chazyan as they seem, they contain fossils that have been considered distinctive of the "Marmor", "Ashby" and "Porterfield" stages in the Appalachians, rather than of the first alone.

The faunas from the Youngman and the St. Dominique formations have strong similarities to those of the "Ashby" and "Porterfield" units in Virginia. Among the brachiopods, *Christiania*, *Playtmena* and *Leptellina* are particularly suggestive of "Porterfield", *Valcourea* of "Ashby". Trilobites such as *Lonchodomas* are frequent in argillaceous "Porterfield" rocks. The coral *Billingsaria* is present in small reefs in the St. Dominique, as it is in the Crown Point and Valcour limestones of the type Chazyan, the Lincolnshire of the type "Ashby", and the Lenoir limestone of Tennessee (Twenhofel et al., 1954, p. 274). Ostracodes in the Crown Point are said to be like those in upper "Ashby" (Lincolnshire) and "Porterfield" (Edinburg) (Swain and Kraft, 1956; Swain, 1957). The fossils in the more argillaceous Youngman generally resemble those in the "Porterfield"; those in the St. Dominique, resemble those in the type Chazy and the "Ashby.",

The writer has thought that the Bolarian Series (that is, the "Porterfield" and younger pre-Trentonian) in the Appalachians is younger than the Chazyan, and that its lower part is older than the type Black River (Kay, 1948; 1956). Cooper (1956) thought the Chazy considerably older, in a stage ("Marmor") separated by the "Ashby" from the "Porterfield" (Arline); but with the "Ashby" missing in the section above the type "Marmor" (Lenoir). Rodgers (1953; 1954) thought the Christiania-bearing "Porterfield" (Arline of Copper) a facies of the "Marmor" (Lenoir) rather than younger and separated by a disconformity. The Highgate Springs sequence

strongly suggests that *Christiania*-bearing beds are in a facies of Chazyan and are correlative of "Ashby" and of "Porterfield". Thus the base of the Bolarian Series may lie within the upper Chazyan Series rather than above it.

Classification.—Correlations within the post-Canadian pre-Trentonian rocks cannot be made with assurance between distant sequences. In the north, it is possible by faunal and lithic means to subdivide the Chazyan and Blackriveran series each into three stages. In the Appalachians, it seems best to establish provincial stages from a single sequence such as that in the Virginias; "Blackford", "Ashby" and "Porterfield" are in such a sequence. The place of the "Marmor" stage is in dispute; it is not in the same sequence. The "Porterfield" and the pre-Trentonian part of the "Wilderness" stage constitute the type Bolarian Series. The "Blackford" and "Ashby" stages are believed by the writer to be Chazyan, but not to constitute the whole of that series; they are in a provincial series that is approximately the St. Paul Group (Neuman, 1951). The classifications are summarized in table 3.

TABLE 3

Classification of part of the Ordovician System

NEW YORK		VIRGINIA	
TRENTONIAN SERIES – Rocklandian stage			
BLACKRIVERAN	Chaumontian	L. "Wilderness"	BOLARIAN
	Lowvillian		
	Pamelian	"Porterfield"	
CHAZYAN	Valcourian		
	Crownian	"Ashby"	SAINT PAUL
	Dayan	"Blackford"	
CANADIAN			

CONCLUSIONS

The Highgate Springs sequence forms a very narrow but persistent thrust slice extending for eighty miles across the international boundary from Quebec to Vermont. The sediments range from upper Canadian Beldens limestone to middle Trentonian Hortonville shale. The Chazyan rocks are of particular interest because they change facies from St. Dominique limestone resembling typical Chazyan into Youngman calcareous shales having a sparse but distinctive fauna somewhat unlike that in typical Chazyan, and resembling lower Bolarian and older faunas in the Appalachians. The slice is the attenuated east limb of a synclinorium that seems to have slipped on an horizon in the upper Canadian Series. Submarine breccias in the medial Trentonian

in the synclinorium are of rocks like those in the Highgate Springs slice, suggesting that the deformation was initiated in the Trentonian epoch.

The Chazyan Series seems equivalent to the strata in the Appalachians contained in the Blackford, Ashby and lower Porterfield "stages". The succeeding Black Riveran Series seems to be contained in the Bolarian Series of the Appalachians, but the lowest Bolarian seems older and equivalent to upper Chazyan.

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NATURE OF GLAUCOPHANITIC METAMORPHISM

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ABSTRACT. The compositions, optics, and occurrences of amphiboles of the riebeckite-glaucophane group are reviewed, and their genesis is discussed.

Probably, glaucophane is stable only at high pressures and low temperatures and the presence of riebeckite and magnesioriebeckite molecules in solid solution extends the stability field of glaucophanic amphiboles in pressure and temperature. Jadeite has been proved to be stable at high pressures and low temperatures and the presence of aegirine molecule in solid solution probably extends the stability field of jadeitic pyroxenes in pressure and temperature.

The metamorphism producing glaucophanites may be regarded as representing higher pressures than the Dalradian type of metamorphism. As examples, the glaucophanitic metamorphisms in the Omi district and in the Sanbagawa-Mikabu Zone are described in some detail.

INTRODUCTION

A number of contrasting views on the origin of glaucophane schists have been expressed in geologic literature. The most important of the disputed problems is why glaucophane schists were formed instead of more common metamorphic rocks such as green schists, amphibolites and biotite schists.

One group of geologists considers that glaucophane schists are formed under peculiar chemical conditions during metamorphism, especially richness in soda. For example, Harker (1932) considered that glaucophane schists are formed either by metamorphism of alkalic rocks (especially of the spilite series) or by introduction of soda into sediments. Taliaferro (1943) claimed that metasomatic introduction of soda and certain other components into the country rocks near contact with ultrabasic intrusives is essential for the formation of glaucophane schists.

Some glaucophane schists are indeed rich in soda and such compositions may be due to soda-metasomatism. There exist, however, other glaucophane schists that are not so rich in soda. Washington's (1901) attempt to discover constant chemical differences between glaucophane schists and ordinary amphibolites was unsuccessful. Some glaucophane schists are chemically identical to amphibolites and green schists. They are different because in glaucophane schists the soda has gone into glaucophane, whereas in amphibolites and green schists it has gone into plagioclase. Furthermore, we can make chemical equations between the mineral assemblages of glaucophane schists on the one hand and those of amphibolites and green schists on the other, as shown below.

Therefore, peculiar chemical conditions and soda-metasomatism cannot be indispensable for the formation of glaucophane schists. It follows that glaucophane schists must be formed under a certain range of physical conditions, distinct from that governing the formation of green schists and amphibolites. This view was advanced by Eskola (1929, 1939) and was supported by de Roever (1950, 1955a, 1955b). Eskola proposed the concept of the glaucophane-schist facies, which represents a specific range of physical conditions, and de Roever tried to divide the facies into subfacies. They both considered that the temperature range of the glaucophane-schist facies is essentially the same as that of the green schist and epidote-amphibolite facies,

and that the glaucophane-schist facies is a high pressure equivalent of the latter two facies.

This paper discusses the origin of glaucophane schists with special reference to the stability of the minerals characteristic of glaucophane schists, such as glaucophanic amphiboles and jadeitic pyroxenes. The stability of glaucophanic amphiboles and jadeitic pyroxenes varies not only with temperature and pressure but also with their chemical compositions. Lack of due attention to this fact seems to have been a cause of fruitless discussions. The existence of the glaucophane schist facies does not preclude the effect of chemical factors on the formation of glaucophane schists.

In this paper, the term *glaucophanites* indicates rocks that contain glaucophane as one of the main constituents. Therefore, glaucophanites include glaucophane schists. The metamorphism producing glacophanites will be called *glaucophanitic metamorphism*.

MINERALOGY OF THE RIEBECKITE-GLAUCOPHANE GROUP

Subdivisions and Nomenclature.—We begin with a brief survey of the compositions and modes of occurrence of the riebeckite-glaucophane group of alkali amphiboles. For a more complete discussion of the chemical compositions, optic properties and genesis of the alkali amphiboles, see Miyashiro (1957).

The idealized formula of the riebeckite-glaucophane group is as follows: $Na_2R''_3R'''_2Si_8O_{22}(OH)_2$, where R'' represents divalent atoms (mainly Mg and Fe'') and R''' represents trivalent ones (mainly Al and Fe'''), both in 6-coordination (R''' does not include the Al atoms replacing Si).

Figure 1 (A) shows the degree of the Al←→Fe''' substitution in R''' and that of the Mg←→Fe'' substitution in R'' for reliable analyses of the members of the riebeckite-glaucophane group. The members with Fe'''/R''' ratios lower than 0.7 occur characteristically in glaucophanitic metamorphic rocks, whereas those with Fe'''/R''' ratios higher than 0.7 occur not only in glaucophanitic metamorphic rocks but also in other metamorphic and alkalic igneous rocks. Thus it is in harmony with petrographic custom and convenience to call all the members with Fe'''/R''' ratios lower than 0.7 "*glaucophane*" or more strictly "*glaucophane in a broad sense*".

Figure 1 (B) shows the subdivisions of glaucophane and other members of the riebeckite-glaucophane group. The members rich in Al and Mg (Fe'''/R'''<0.3, Fe''/R''<0.5) are called *glaucophane proper*, and their ferrous analogues are called *ferroglaucophane*. The members with intermediate values of Fe'''/R''' ratio (0.3-0.7) are called *subglaucophane*. The members rich in Fe''' and Fe''' (Fe'''/R'''>0.7, Fe''/R''>0.5) are called *riebeckite*, and their magnesium analogues are called *magnesioriebeckite*.

The idealized formulas of the main members of the riebeckite-glaucophane group are as follows:

Riebeckite	$Na_2Fe_3''Fe_2'''Si_8O_{22}(OH)_2$
Magnesioriebeckite	$Na_2Mg_3Fe_2'''Si_8O_{22}(OH)_2$
Glaucophane proper	$Na_2Mg_3Al_2Si_8O_{22}(OH)_2$

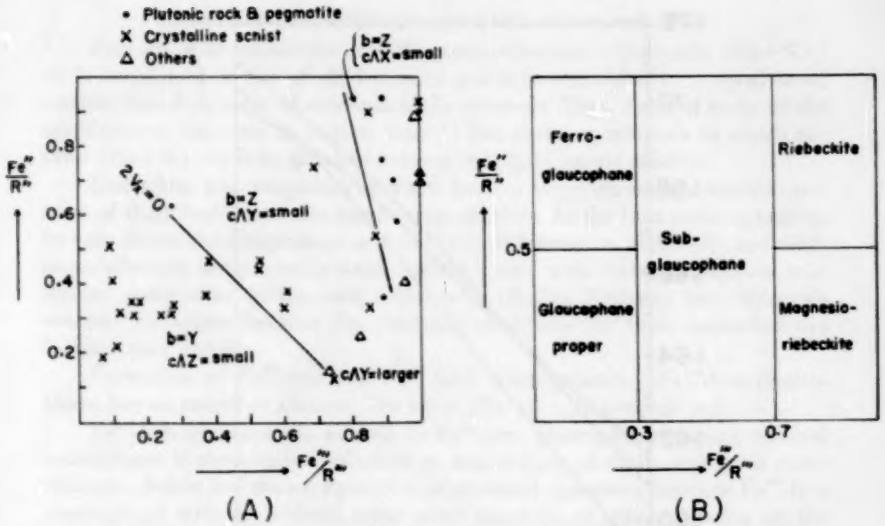

Fig. 1. The Al-Fe‴ and Mg-Fe″ substitutions in the riebeckite-glaucophane group. (A) Modes of occurrence and optic orientation, (B) Nomenclature.

Optic Properties.—The refractive indices increase with the increase of Fe‴ and Fe″, that is, from the corner of glaucophane proper diagonally to that of riebeckite. The index β varies from 1.61 to 1.70, as shown in figure 2.

The variation of the optic orientation with composition is shown in figure 1 (A). In glaucophane rich in Al and Mg, b=Y, cΛZ=small (4°-15°), and $2V_x$=50°-small. Such glaucophane may be called *parallel-symmetric glaucophane*, because the optic plane is parallel to (010). (Certain writers confine the term glaucophane to this parallel-symmetric glaucophane, but this usage would lead to undesirable confusion.) In chemical composition, parallel-symmetric glaucophanes belong either to glaucophane proper or to Mg-rich members of subglaucophane.

With the increase of Fe‴ and Fe″, the optic angle becomes smaller down to zero. With a further increase, it becomes larger again in the optic plane normal to (010). Thus, b=Z and cΛY=small (2°-15°). Such a normal-symmetric amphibole has been usually called *crossite*. In chemical composition, crossites belong mostly to subglaucophane, rarely to magnesioriebeckite and riebeckite. However, the accurate position of the field boundary between parallel-symmetric glaucophane and crossite has not been established.

In riebeckite rich in Fe‴ and Fe″, b=Z and cΛX=small (0°-5°). The optic properties of the transitional minerals between crossite and riebeckite with cΛX=small are not clear. Figure 2 shows a possible interpretation.

In the amphiboles near the diagonal from glaucophane proper to riebeckite, the extinction angle is smaller than 15°, as mentioned above. However, in magnesioriebeckites with very low Fe″/R″ ratios, the extinction angle is much larger, reaching about 30°, with very strong dispersion. A remarkable

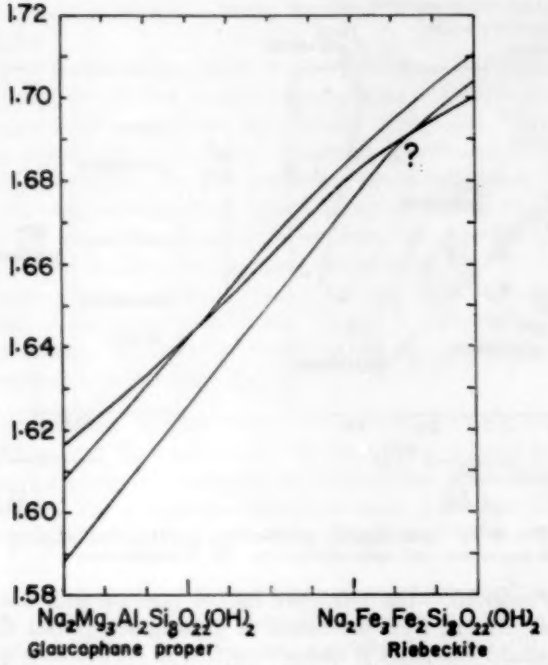

Fig. 2. Increases in refractive indices along the diagonal from glaucophane proper to riebeckite.

feature of such amphiboles is that two of the principal axes of the absorption ellipsoid show a large departure from the axes of the indicatrix. One of the absorption axes is parallel to the b axis, another is practically parallel to the c axis, and the remaining is practically normal to both b and c. α=1.65-1.66 and γ-α=about 0.01. (Amphiboles characterized by such peculiar optic properties have been described under the name of torendrikite by certain writers.)

Modes of Occurrence.—Riebeckite occurs in alkalic igneous rocks, in crystalline schists, and in metamorphosed ironstones (as crocidolite).

Magnesioriebeckite occurs rarely in alkalic igneous rocks, but more commonly in crystalline schists and other metamorphic rocks (sometimes as crocidolite).

Metamorphic riebeckites and magnesioriebeckites generally have larger R''' values than igneous ones. Miyashiro (1957, Chapters VIII and X) has shown that the compositions of such alkali amphiboles vary with their formation temperature.

Subglaucophane and glaucophane proper occur only in crystalline schists, and such associated rocks as veins cutting through schists.

Ferroglaucophane with relatively low Fe''/R'' ratios occurs in crystalline schists, but that with relatively high Fe''/R'' ratios has not been found in any rock, so far as we are aware.

GENESIS OF GLAUCOPHANE AND SOME OTHER MINERALS

Formation of Riebeckite and Magnesioriebeckite.—The ratio (Na+K)/Al is equal to 1 in the alkali feldspars and 0 in anorthite. It is equal to or smaller than 1 in most of common mafic minerals. Thus, in most rocks of the earth's crust, the ratio is smaller than 1. The rocks or minerals in which the ratio (Na+K)/Al is larger than 1 are said to have *excess alkalies.*

Riebeckite and magnesioriebeckite have a large excess of alkalies, and most of their host rocks also have excess alkalies. As the host rocks appear to be very diverse in temperature and pressure of formation, riebeckite and magnesioriebeckite are probably stable within a very wide range of physical conditions under some environment with excess alkalies. Probably these minerals are rare in nature because the chemical conditions for their formation are fulfilled only rarely.

Formation of Fe'''-free and Fe'''-poor Glaucophanes.—Fe'''-free glaucophane has no excess of alkalies. The ratio (Na+K)/Al is equal to 1.

Fe'''-free glaucophane as well as Fe'''-free glaucophane-bearing mineral assemblages is chemically equivalent to assemblages of albite and some mafic minerals. Below are shown examples of chemical equations between Fe'''-free glaucophane with or without some other minerals of glaucophanites on the one hand and some assemblages of green schists on the other.

$$\underset{\text{albite}}{2\ NaAlSi_3O_8} + \underset{\substack{\text{antigorite mol.}\\ \text{(in chlorite)}}}{H_4Mg_3Si_2O_9} = \underset{\text{glaucophane}}{H_2Na_2Mg_3Al_2Si_8O_{24}} + H_2O$$

$$\underset{\text{albite}}{50\ NaAlSi_3O_8} + \underset{\text{chlorite}}{9\ H_8Mg_5Al_2Si_3O_{18}} + \underset{\text{actinolite}}{6\ H_2Ca_2Mg_5Si_8O_{24}}$$

$$= \underset{\text{glaucophane}}{25\ H_2Na_2Mg_3Al_2Si_8O_{24}} + \underset{\text{epidote}}{6\ HCa_2Al_3Si_3O_{13}} + \underset{\text{quartz}}{7\ SiO_2} + 14\ H_2O$$

$$\underset{\text{albite}}{10\ NaAlSi_3O_8} + \underset{\text{chlorite}}{3\ H_8Mg_5Al_2Si_3O_{18}} + \underset{\text{epidote}}{6\ HCa_2Al_3Si_3O_{13}} + \underset{\text{quartz}}{7\ SiO_2} + 14\ H_2O$$

$$= \underset{\text{glaucophane}}{5\ H_2Na_2Mg_3Al_2Si_8O_{24}} + \underset{\text{lawsonite}}{12\ H_4CaAl_2Si_2O_{10}}$$

From these equations it is clear that the chemical conditions for the formation of Fe'''-free glaucophane are fulfilled in most basic rocks and some sediments. Fe'''-free glaucophane can be formed in rocks of common compositions without excess alkalies only if the physical conditions prevailing are within an appropriate range. Glaucophane proper and ferroglaucophane resemble Fe'''-free glaucophane in composition, and so can probably be formed in many basic rocks and some sediments metamorphosed under certain physical conditions.

The principles governing osmotic equilibrium in geological problems were discussed in detail by Thompson (1955). In the above equations, the total volume of the solid phases on the glaucophane-bearing side is much smaller than that on the other side. Thus, the glaucophane-bearing side represents higher solid pressures than the other, if the temperature and water pressure be constant.

Sliding Equilibrium in the Riebeckite-Glaucophane Group.—Subglaucophane is intermediate in composition between Fe'''-free glaucophane and

riebeckite-magnesioriebeckite. Thus, its formation is probably controlled by high solid pressure as well as by the supply of excess alkalies. It is to be expected that sliding equilibria should hold between riebeckite-glaucophane solid solutions and minerals of green schists or epidote-amphibolites under some conditions. The general problems of sliding equilibria in rock-forming minerals have been discussed in detail by Ramberg (1945).

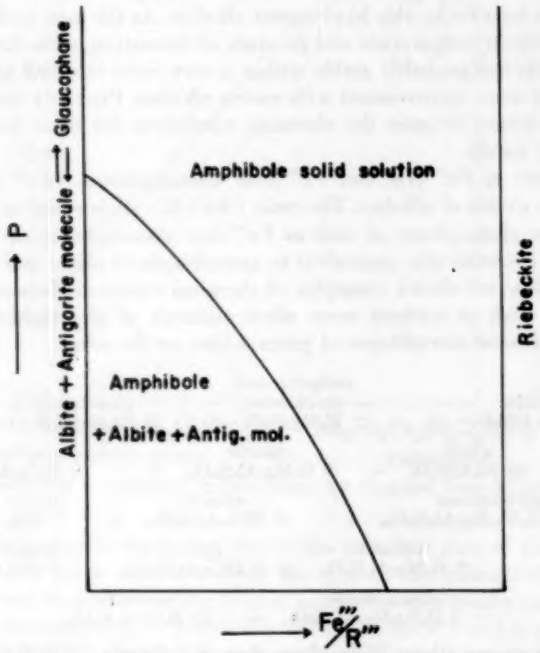

Fig. 3. Hypothetical diagram showing a possible sliding equilibrium between alkali amphibole and albite + antigorite molecule in chlorite.

Figure 3 shows a sliding equilibrium with regard to solid pressure, corresponding to the first of the equations. Probably, the composition range of amphiboles of the riebeckite-glaucophane group becomes generally larger toward the lower Fe'''/R''' values with increasing solid pressure. The curve of this figure would shift upwards or downwards with variation in temperature and by the formation of solid solution in the minerals concerned.

Zonal Mapping of Glaucophanitic Metamorphic Terrains.—It is rather surprising that zonal mapping of glaucophanite-bearing metamorphic terrains on the basis of progressive mineralogical changes had not been carried out anywhere before one of us succeeded in it in 1956 in his study of the schists of the Omi district in Japan (Banno, 1956, 1958). He distinguished the zone of chlorite and that of biotite, and found that glaucophane is confined to the former, that is, to the lower grade.

Immediately afterwards, Yôtarô Seki of Saitama University (personal communication) also succeeded in zonal mapping in the glaucophanitic metamorphic terrain of the Kanto Mountains belonging to the Sanbagawa-Mikabu Zone. According to him, glaucophane is confined to the lower-grade part of the terrain, lawsonite is confined to a very limited area within the glaucophane-bearing part, and pumpellyite occurs in the same and adjacent areas, whereas epidote occurs widely in almost all parts of the terrain.

Still later, Banno found that in the Bessi district of the Sanbagawa-Mikabu Zone glaucophane is confined to the lower-grade part.

Thus, the greater stability of glaucophane in the lower grade, that is, probably at lower temperatures, has been established. (The glaucophanite-bearing zone was found to grade directly into unmetamorphosed rocks in the Kanto Mountains and elsewhere.)

The effect of decreasing temperature on the stability of glaucophane is probably similar to that of increasing pressure. The composition range of amphiboles of the riebeckite-glaucophane group should become larger toward the lower Fe'''/R''' values with decreasing temperature. This relation has been noticed to hold good in the Bessi district.

Supply of Excess Alkalies.—In glaucophane proper, ferroglaucophane, and contiguous amphiboles, the amount of excess alkalies (i.e. the riebeckite and magnesioriebeckite molecules) is small. Probably such a small amount can readily be supplied to the minerals by some adjustments of the associated minerals without any introduction of materials from outside, for fictive reactions to produce excess alkalies from albite, involving decreases of volume in solid phases, would proceed under high solid pressures. Probably metasomatic introduction of alkalies is not necessary for the formation of glaucophane proper and ferroglaucophane, though this statement does not preclude the possibility of such introduction in some cases.

Riebeckite, magnesioriebeckite and contiguous amphiboles have larger amounts of excess alkalies, and many of their host rocks also have excess alkalies. Schists with much excess alkalies may contain not only riebeckite but also aegirine, as was described by Suzuki (1934). Since rocks with excess alkalies are rare in the earth's crust, such metamorphic rocks may have been formed in most cases by metasomatic introduction of alkalies from outside.

However, in some cases, rocks with excess alkalies would be formed by metamorphic differentiation. Miyashiro and Iwasaki (1957) described an example of magnesioriebeckite and aegirine formed in an alkali-deficient schist with $(Na+K)/Al=0.66$ owing to adjustments of the associated minerals probably under high solid pressure combined with low temperature. Under such conditions metamorphic differentiation may produce rocks that are rich in such amphibole and aegirine, thus leading to the formation of rocks with excess alkalies in bulk composition.

Stability of Jadeite and Aegirine.—Robertson, Birch and MacDonald (1957) clarified the stability field of jadeite. At room temperature, the formation of jadeite from albite and nepheline requires a solid pressure of about 1400 kg/sq.cm, and at higher temperatures the required pressure is higher. In

the environment with excess silica, the pressure required for the formation of jadeite from albite is much higher.

Thus, jadeite and contiguous pyroxenes are high-pressure and low-temperature minerals. On the other hand, aegirine and contiguous pyroxenes are stable probably within a very wide range of pressure and temperature in some environments with excess alkalies. Accordingly, the relation between jadeite and aegirine closely resembles that between glaucophane and riebeckite. Probably, the presence of aegirine molecule in solid solution stabilizes jadeitic pyroxenes at a larger range of pressure and temperatures. The sliding equilibrium between jadeite-aegirine solid solution and albite with regard to solid pressure is probably as shown in figure 4.

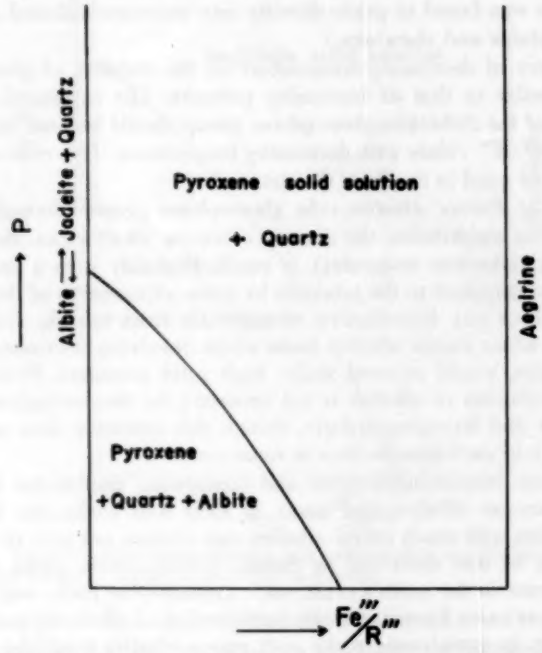

Fig. 4. Hypothetical diagram showing a possible sliding equilibrium between alkali-pyroxene + quartz and albite.

The genetical relationship of jadeite and contiguous pyroxenes with glaucophanitic metamorphism has been recently clarified by de Roever (1947, 1950, 1955b) in Celebes and by Bloxam (1956) in California. These minerals occur in low-grade parts of typical glaucophanitic metamorphism, indicating that the metamorphism represents very high pressures. On the other hand, aegirine and contiguous pyroxenes may occur in any grades of any type of metamorphism, but only if the chemical conditions are appropriate.

Stability of Epidote, Pumpellyite and Lawsonite.—Epidote, pumpellyite and lawsonite are dense hydrous silicates resembling anorthite in composition.

Epidote is formed widely in the glaucophanitic as well as in Dalradian type of metamorphism, whereas lawsonite is characteristic of the glaucophanitic type. Pumpellyite is lacking in the Dalradian metamorphism, and occurs in some glaucophane-bearing and glaucophane-free metamorphic areas.

The three minerals are somewhat different in composition from one another. Consequently, their stability relations are controlled not only by physical but also by chemical conditions. Under some conditions, two of them or even all three may coexist in stable equilibrium.

Reactions to form lawsonite from green schist minerals may be represented by the following equations:

$$\underset{\text{epidote}}{2\ HCa_2Al_3Si_3O_{13}} + \underset{\text{kaolinite}}{H_4Al_2Si_2O_9} + 5\ H_2O = \underset{\text{lawsonite}}{4\ H_4CaAl_2Si_2O_{10}}$$

$$\underset{\text{albite}}{10\ NaAlSi_3O_8} + \underset{\text{chlorite}}{3\ H_8Mg_5Al_2Si_3O_{18}} + \underset{\text{epidote}}{6\ HCa_2Al_3Si_3O_{13}} + \underset{\text{quartz}}{7\ SiO_2} + 14\ H_2O$$

$$= \underset{\text{glaucophane}}{5\ H_2Na_2Mg_3Al_2Si_8O_{24}} + \underset{\text{lawsonite}}{12\ H_4CaAl_2Si_2O_{10}}$$

The total volume of the solid phases increases in the first equation and decreases in the second. High water pressure favors the formation of lawsonite and pumpellyite. The occurrence of these minerals in some glaucophanitic metamorphic rocks may suggest that the metamorphism takes place under high water pressures at low temperatures.

Biotite and Pyralspite Garnet.—The relative positions of various isograds differ in different types of metamorphism, and consequently may be used to distinguish the types. For example, in the Dalradian type, the biotite isograd is in a lower grade than the almandine isograd, whereas in some glaucophanitic metamorphism the biotite isograd practically coincides with the almandine isograd, and in others biotite does not occur though almandine does occur commonly. Both biotite and almandine isograds should shift with variations in solid and water pressures. The change in the relative position of the biotite and almandine isograds is a result of their shifting in different manners.

Miyashiro (1953) has shown that the Mn content of pyralspite garnet in ordinary pelitic metamorphic rocks generally decreases with increasing metamorphic grade, but the absolute value of the Mn content differs in different types of metamorphism, being lower in metamorphism under higher pressures. This is one of the factors governing the shifting of the almandine isograd. However, the occurrence of almandine is related also to the compositions of the associated minerals.

SCHISTS OF THE OMI DISTRICT IN JAPAN

General Statement.—The Omi schists are exposed in an area about 5 km in width and 10 km in length in Omi-mati (town of Omi) and its vicinity of Niigata Prefecture (about 137° 45′ E, 36° 57′ N). The schists are in fault contact with Permo-Carboniferous and Lower Jurassic formations. Serpentines are intruded along the boundary faults and also into the schists and Permo-Carboniferous. As closely related schists in the adjacent area are covered unconformably by the Lower Jurassic, the age of the metamorphism must be pre-Jurassic.

Pebbles of serpentine were found in the conglomerate of the Jurassic. Probably all the serpentines are genetically related to the schists and were intruded during the metamorphism of pre-Jurassic time. Some of the serpentines contain jadeite masses. The serpentines along the boundary faults are considered to have been emplaced in the solid state during the post-Jurassic fault movement.

Many years ago, jadeite was found to occur in Kotaki (Kawano, 1939; Omori, 1939; Yoder, 1950; Iwao, 1953). Kotaki is about 4 km to the southeast of the present schist area. There the jadeite occurs in albitite masses within serpentine. The serpentine belongs to the same group as the serpentines of the present area.

The grade of metamorphism increases gradually from the boundary faults toward the central part of the schist area where a big mass of serpentine is exposed. The schist area is divided into the zones of chlorite and of biotite, the latter representing a higher grade than the former. The rocks of the

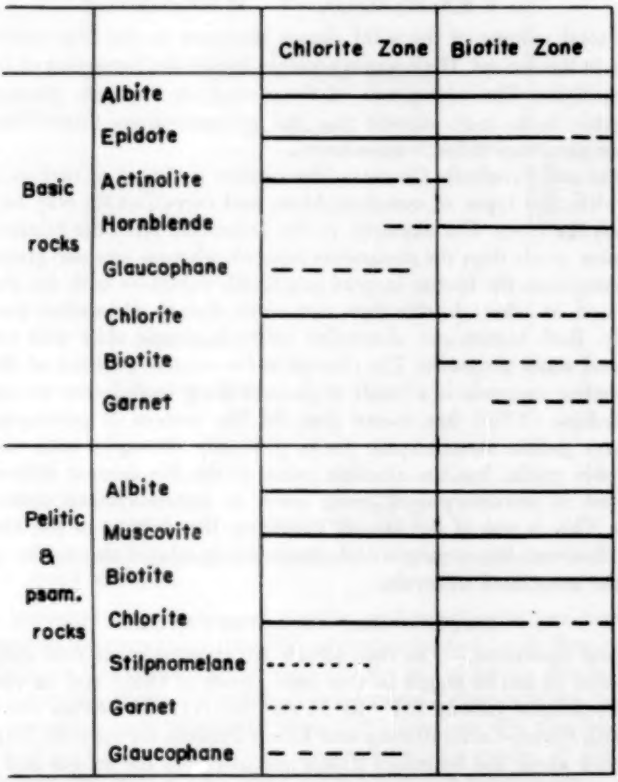

Fig. 5. Mineralogical variations with increasing metamorphism in the Omi schists. A full line indicates that the mineral concerned is common, a broken line indicates that it is less common, and a dotted line indicates that it is very rare.

chlorite zone are devoid of biotite and generally devoid of pyralspite garnet, whereas most of the pelitic rocks of the biotite zone contain not only biotite but also almandine. Therefore, the biotite isograd practically coincides with the almandine isograd. Glaucophane is confined to the chlorite zone.

The relation between mineral composition and metamorphic grade is shown in figure 5.

Chlorite Zone.—The most common type of pelitic rock in this zone is graphite-calcite-chlorite-sericite-albite-quartz schist. The most common type of basic rock is actinolite-chlorite-epidote-albite schist with small amounts of sericite, rutile, sphene and calcite. In the neighborhood of the biotite isograd, actinolite-bearing rocks occur interbedded with rocks having blue-green common hornblende. Some rocks of this zone contain piedmontite or stilpnomelane. (Stilpnomelane was identified by the x-ray powder method.)

Glaucophane-bearing rocks occur rather rarely and in association with glaucophane-free rocks. They do not show any relationship to the serpentines, nor any evidence of soda-metasomatism. Probably some of them were derived from basic materials and others from pelitic rocks mixed with some basic materials. The following rock types were noticed:

(a) Massive chlorite-albite-glaucophane-epidote rock with or without actinolite.
(b) Sericite-chlorite-epidote-albite-glaucophane schist.
(c) Sericite-chlorite-garnet-epidote-glaucophane-albite-quartz schist.
(d) Muscovite-epidote-garnet-glaucophane-quartz schist.

Sphene, rutile, and tourmaline occur sparingly in some rocks.

Some of the glaucophanes are parallel-symmetric with $c\Lambda Z = 8°\text{-}15°$, whereas the others are normal-symmetric with $c\Lambda Y = 15°\text{-}20°$. The optic angle about X ranges from 0° to 56°. The index β ranges from 1.646 to 1.664. These optic properties suggest that the glaucophanes belong to subglaucophane. In rock type (a), some of the glaucophanes are rimmed by actinolite, and in some of the others actinolites are rimmed by glaucophane.

The analysis of a garnet from a rock of type (d) is as follows: SiO_2 38.25, TiO_2 0.52, Al_2O_3 19.57, Fe_2O_3 3.02, FeO 21.12, MnO 5.48, MgO 2.31, CaO 7.56, H_2O_+ 2.48, H_2O_- 0.07, P_2O_5 0.15, total = 100.53; a = 11.62 A.

Biotite Zone.—The most common type of pelitic rock in this zone is almandine-chlorite-muscovite-biotite-albite-quartz schist. Hornblende occurs in some pelitic rocks. Graphite, sphene, rutile, apatite, and tourmaline occur in small amounts. Chlorite disappears in the highest grade.

The most common type of basic rock is albite-epidote-hornblende schist, usually with a small amount of chlorite. The hornblende is common hornblende whose axial color of Z is blue-green. Chlorite disappears in the highest grade. Some basic rocks contain quartz, biotite and/or almandine. Sphene, rutile, calcite and opaque minerals occur in small amounts. The anorthite content of the "albite" is 12 percent or less.

Additional Notes.—Neither jadeite nor aegirine has been found in the schists themselves, though jadeite occurs in the associated serpentines.

Lawsonite and pumpellyite have not been found. For a more detailed description of the Omi schists, see Banno (1958).

SCHISTS OF THE SANBAGAWA-MIKABU ZONE

General Statement.—The Sanbagawa-Mikabu Zone is a long belt through the Japanese Islands, and its metamorphism probably differs in character to some extent in different parts. Suzuki (1930) made a systematic description of the rocks of the Zone in Sikoku. Except in the Bessi and adjacent districts, the metamorphism did not reach the grade at which biotite is formed. Glaucophane-bearing rocks occur intermingled with actinolite-green schist, etc. in many parts of the Zone.

Most of the glaucophanes of the Sanbagawa-Mikabu Zone are normal-symmetric with $\beta = 1.645$-1.669 and $c\Lambda Y = 16°$-$23°$ (Horikosi, 1936). The high refractive indices and large extinction angles suggest that they belong to subglaucophane near magnesioriebeckite, and not to glaucophane proper. Magnesioriebeckite in the strict sense also occurs (Miyashiro and Iwasaki, 1957). Aegirine-rich pyroxenes were found very rarely, but jadeite has not been found. Epidote is stable in most, if not all, grades.

In the Kanto Mountains glaucophane is confined to the lower grade part, lawsonite and pumpellyite occur in very limited areas within the same part, and epidote occurs in almost all grades including the lawsonite-and/or pumpellyite-bearing areas (Seki, 1957a,b and personal communication).

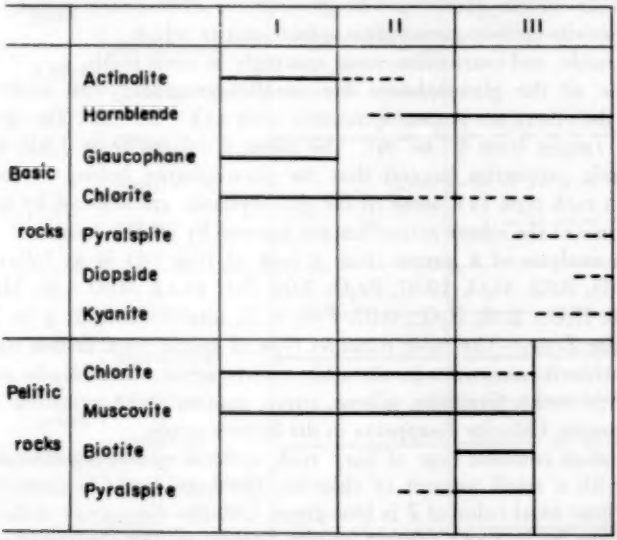

Fig. 6. Mineralogical variations with increasing metamorphism in the Bessi district. A full line indicates that the mineral concerned is common; a broken line indicates that it is rare. Epidote is stable in all the grades. The plagioclase is albite in zones I and II and in the lower-grade part of zone III, whereas it is sodic oligoclase in the high-grade part of zone III. The porphyroblastic development of plagioclase is conspicuous in the grades higher than the middle of zone II.

Bessi District.—Figure 6 shows the mineralogical changes with increasing grade of metamorphism in the Bessi District of the Zone, investigated by Banno. The metamorphic area is divided into three zones representing progressive changes as follows:

(a) Zone I is characterized by the occurrence of glaucophane in some basic rocks. The glaucophanes belong to crossite in most cases at least. In the lower-grade part of this zone, recrystallization is very incomplete.

(b) Zone II is characterized by the absence of glaucophane and of biotite. Though glaucophane is confined to the preceding zone, magnesioriebeckite occurs in all three zones.

(c) Zone III is characterized by the occurrence of biotite in pelitic rocks. The higher-grade part of this zone is represented by a big mass of amphibolite, and so does not have pelitic rocks.

Epidote is stable in all the grades, pumpellyite occurs in a lower-grade part of zone I, and lawsonite has not been found in this district.

Even in zone I, glaucophane-bearing rocks are much less common than glaucophane-free ones.

ACKNOWLEDGMENTS

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STATISTICAL STUDY OF TOPOGRAPHY, SHEETING, AND JOINTING IN GRANITE, ACADIA NATIONAL PARK, MAINE

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ABSTRACT. Both field and laboratory data are used to show the relationship between topography and structure in a mountainous area of subaerial erosion that was reshaped by glacial ice. A statistical method is employed to assist in relating effects of stream and glacial erosion, as controlled by sheeting and jointing, upon homogeneous granite. This method allows one to depict a topographic surface, in terms of slope directions and inclinations, by a simple diagram similar to petrofabric-type diagrams.

Two such diagrams representing the present topography are prepared, one from map data and one from field data. Another diagram based on sheeting in the granite is shown to represent rather accurately the preglacial topography. A comparison of preglacial and postglacial topographies indicates three principal changes due to ice action: (1) a shift in main valley trend from N10°E-S10°W to N15°W-S15°E, (2) an increase in easterly and westerly slope areas and slope angles, and (3) the formation of asymmetrical ridge-and-valley cross sections. Additional diagrams of selected portions of the slope profile are employed both to evaluate the method and to further the topographic analysis.

Data on jointing of the granite, when plotted in a rosette diagram, reveal three principal and one subordinate joint sets. The control of each set upon topography is clearly brought out in the analysis.

In terms of the symmetry concept, the ice advance was oblique to the topographic symmetry plane and glacial erosion was adequate to transform the monoclinic pattern to one of triclinic symmetry.

INTRODUCTION

This study is an attempt to interrelate jointing, sheeting, and the topographic effects of stream and glacial erosion. A relatively new statistical method of analysis is employed with the hope of determining its effectiveness as a quantitative as well as qualitative research tool. The area selected for this study represents a small physiographic unit within Acadia National Park in the southeastern part of Mount Desert Island, Maine (fig. 1). It is underlain by petrographically uniform medium-grained granite, and is believed to constitute an ideal area for this type of investigation.

Field studies were made principally by C. A. Chapman and J. P. Wehrenberg during the summer of 1950, and were supported by a generous grant from the University Research Board, University of Illinois. T. T. Quirke, Jr. (1951) and R. L. Rioux (1955) continued the investigation in the laboratory. Subsequent field and laboratory work by C. A. Chapman required some modification of these earlier studies, and the significant findings in all the above investigations are embodied in this paper. Professors G. W. White, P. R. Shaffer, and W. M. Merrill have offered many helpful suggestions. In addition to all above, the writers are greatly indebted to various members of the National Park Service for their cooperation and assistance in connection with this project.

TOPOGRAPHY

Mount Desert Island is an exceptional feature in this region of Maine primarily because of its high relief and rugged profile. A large part of the island is dominated by the Mount Desert Range which trends N 65° E (fig.1). This range is dissected into a series of peaks and ridges by U-shaped

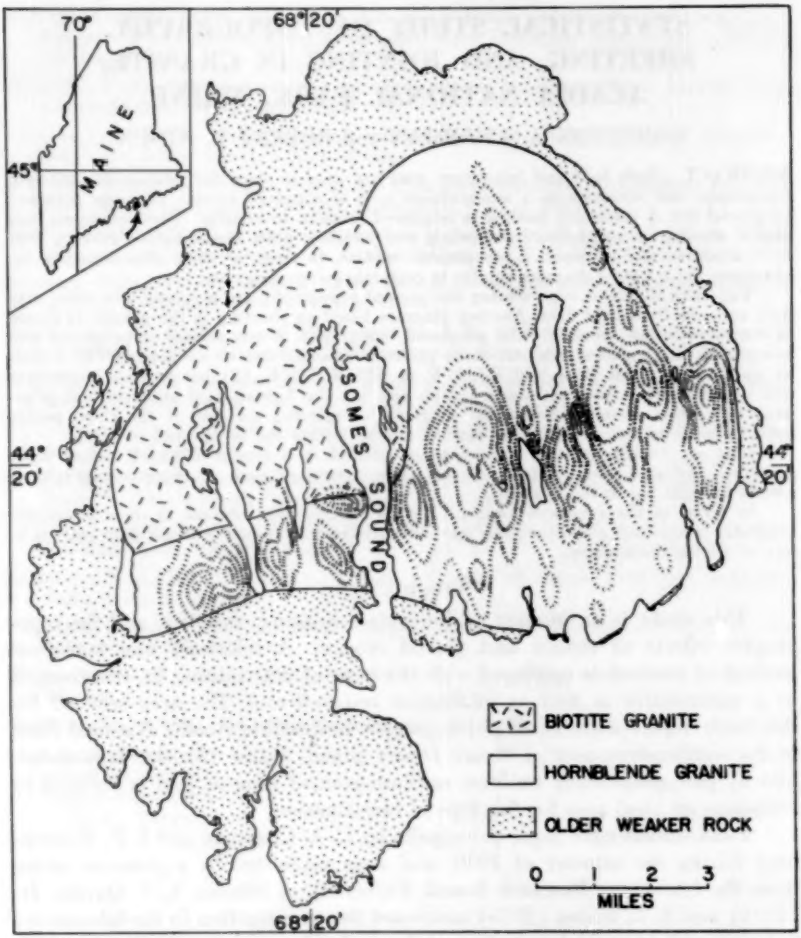

Fig. 1. Index map and simplified relief and geologic map of Mount Desert Island. Acadia National Park includes, in part, most of the area of hornblende granite.

valleys which trend approximately N 15° W (fig. 2). Four of these peaks rise above 1100 feet, and Cadillac Mountain (1530 feet) is the highest point within 20 miles of the Atlantic coastline of the United States (Wood, 1924). Raisz (1929) relates the extensive lowland of the island to the New England peneplane, and he considers the mountain range a monadnock on that erosional surface.

Shaler (1889), Bascom (1919), and Raisz (1929) treat in detail the various physiographic features, but Raisz (1929) gives the most modern and complete account of the physiography and reviews the previous literature. All these earlier workers noted the presence of a dominant master joint set trend-

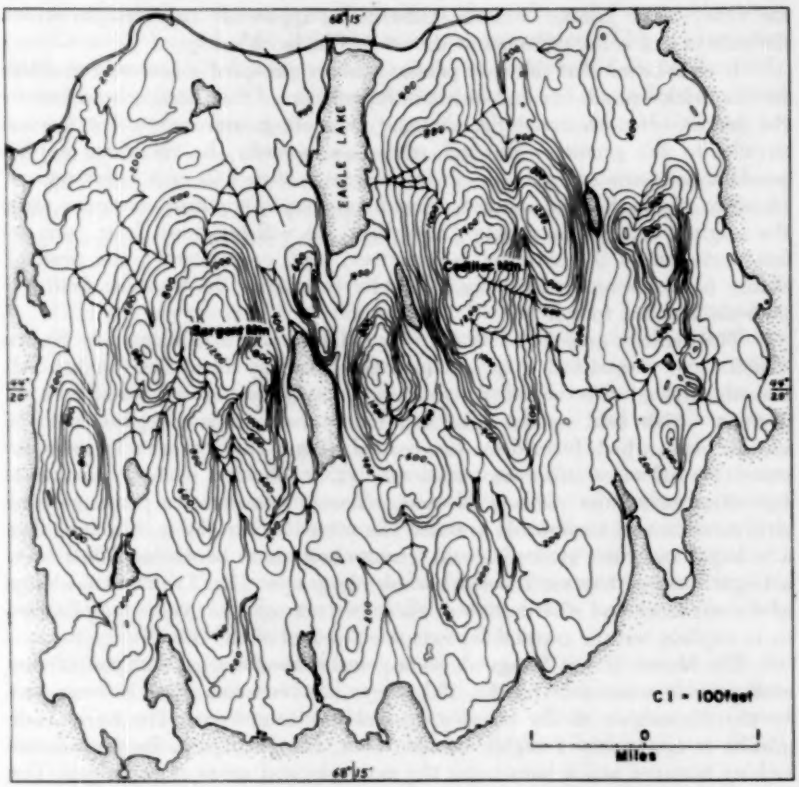

Fig. 2. Topography of southeastern part of Mount Desert Island modified from the United States Geological Survey topographic map of Acadia National Park.

ing roughly N-S and a promient diabase dike set with a similar trend. Since these two features approximately parallel the dominant valleys of the region, these authors attributed the valley trend to the structural control of either joints, or dikes, or both. Chadwick (1939, 1944), on the other hand, concluded that the valleys were controlled by the positions of roof pendants and inclusions of weaker rock and by keystone faults.

No systematic study of jointing and diking had been made of this area, and the above conclusions are based on very general observations. Detailed field work for the present study, however, does show a definite relation between topography and jointing, but three very prominent joint sets were discovered to complicate the interrelation. On the contrary, field observations do not substantiate the earlier conclusion of dike control on topography. For example, basic dikes are as abundant on high ridges as in valleys. A special effort was made to test the roof pendant and inclusion theory, but no evidence in its favor could be found. Contrary to the statements of Chadwick (1939, 1944), inclusions in the granite were not observed to be more abundant in

the valleys. The granite contact, furthermore, apparently cuts straight across the valleys and is not deflected as shown on Chadwick's map.

It is believed that the roof pendant theory occupied a foremost position in Chadwick's mind, because of his interpretation of the sheeting structure in the granite. He attributed the sheeting to cooling, and believed it formed parallel to the granite walls and roof. Consequently, he visualized a roof pendant or downward extension of the granite roof rock in each valley region. In other words, he felt that the present topography virtually conforms with the original roof of the granite "bathylith". As will be shown later, such an interpretation of the sheeting is incorrect; and, consequently, the pendant theory is unsupported. Evidence either for or against the keystone fault hypothesis was not found.

The bedrock geology of Mount Desert Island was first mapped by Shaler (1889) and a modified map was later published by Chadwick (1939). All investigators of this island recognized the existence of a central core of granite, which held up the hills and mountains, and a marginal zone of weaker rocks which formed the low coastal fringe. The fringe rocks are composed of volcanics, siliceous sediments, gabbro-diorite, gneiss, and schist. Intruding these older rocks is the core, generally considered a petrologic and structural unit of hornblende granite. The present study shows, however, that a younger and finer grained biotite granite body is to be distinguished from a larger mass of coarser grained hornblende granite (fig. 1). The knowledge of the existence and distribution of these two petrographic types now enables us to explain certain unusual topographic features of the island.

The Mount Desert Range shows a gourd-shaped outline with the narrow neck pointed westward (fig. 1). This shape has been controlled in large part by the disposition of the hornblende and biotite granites. The hornblende granite is apparently a highly durable rock, due perhaps to its more interlocking texture; and it constitutes the more elevated areas of the island. The biotite granite, on the other hand, has a sugary texture which probably accounts for its high susceptability to weathering and tendency to underlie low ground.

The crestal zone of the mountain mass has an overall trend of approximately N65°E. West of Somes Sound, in the neck of the gourd, the range runs about N70°E coincident with the narrow westward prolongation of the hornblende granite. Though highly symmetrical, this portion of the range shows slightly steeper southerly than northerly slopes.

East of Somes Sound the crestal zone trends about N80°E, and appears to be offset to the north about one mile. This apparent offset in the region of Somes Sound is believed due to the disposition of the two granite masses and not to faulting. Had the biotite granite never intruded, the crestal zone would have extended westward across Somes Sound without apparent offset to the south. The crestal zone east of Somes Sound occupies a rather normal position, that is, about half way between the northern and southern limits of the granite mass. Long gentle ridges trail off to the southern contact, and poorly defined ridges trail off to the northern contact. These latter are generally steep in their higher elevations, and west of Eagle Lake they are interrupted by an

extensive lowland which isolates McFarland Mountain and Youngs Mountain (northwest of Eagle Lake) from the main mountain mass. This lowland was developed by westerly flowing streams which cut back readily across the easily eroded biotite granite. This headward extension of stream valleys eastward has created a pronounced asymmetry to the main mountain range and developed extensive areas of steep, generally northerly slope.

It seems evident, therefore, that the general outline of the Mount Desert Range was controlled by the disposition of a more resistant hornblende granite relative to a weaker granite and the older fringing rocks.

GLACIATION

Evidence of glaciation is abundant in the area. The large transverse valleys are typically U-shaped and many contain rock basin lakes. Boulders up to twenty feet across are widely distributed and a number are precariously perched on the mountain slopes. Some of the largest of these erratics are composed of a coarse granitic rock quite foreign to the island and must have been transported at least 20 to 25 miles.

Shaler (1889) and Raisz (1929) have studied the glacial striations, grooves, and roches moutonnées. They concluded that ice within the valleys moved parallel to present valley trends; whereas ice in the upper levels, which covered the mountain range, moved in a more southeasterly direction. Field work upon which the present paper is based corroborates these earlier findings. Raisz (1929) in the course of field studies was unable to find any supporting evidence for multiple glaciation in this area.

It is agreed by numerous investigators that prior to Pleistocene time the hornblende granite mass formed an elongate range trending roughly perpendicular to the main advance of the ice sheet. The effects of glaciation upon this monadnock are briefly but most admirably described by Raisz (1929, p. 141).

> . . . We have to imagine the mountains at this time as a maturely dissected range with an uneven crest and with deep valleys and dividing spurs on both flanks. The ice ascended the valleys of the northern flank and spilled over the crest, at first, of course, in the saddles, whence long icy tongues descended upon the southern lowlands.
>
> For a long time the ice carved its spillways deeper and deeper into the saddles until it had cut sufficiently deep for more effective discharge, despite the increasing thickness of the ice. Such is the origin of the deep glacial troughs which give so unusual an appearance to the Mount Desert range. Only when the thickness of the ice exceeded the height of the mountains were the crests buried under a slowly moving ice sheet several thousand feet thick.
>
> At this stage the ice movement changed its direction. This is recorded in the asymmetry of the mountains and in the form of the ice-carved ledges. In the beginning, the ice was guided by the direction of the valleys, which was transverse to the main crest and coincided with the direction of the north-south dikes. When the ice extended above the mountain tops, it could not be controlled further by the topography, but followed its own general direction of flow, which was more to the southeast.

THE SSO DIAGRAM

The analysis which follows is based upon a statistical method devised by Chapman (1951 and 1952), and involves the preparation and comparison of statistical slope orientation diagrams (SSO diagrams). For each diagram the direction and inclination of slope are measured at several hundred uniformly

spaced points on a topographic map to obtain a representative sample of the orientation of the various slopes of an area. For this study the 1942 edition of the Mount Desert Island, Maine topographic map with a scale of 1:31680 and a contour interval of twenty feet was used. Points were selected by means of a square grid the spacings of which were governed roughly by the number of points desired. Slope measurements were determined at each grid intersection on the map by means of a protractor and slope scale. Each measurement, thus, represents the orientation in space of a plane tangent to the topographic surface at the point of measurement.

Using what is essentially equivalent to radial coordinate paper, with the origin at the center and north established at the top of the sheet, each set of slope measurements may be plotted as a point. The angle of slope (inclination) is indicated by the distance of the point from the origin or plot center. This distance increases as slope angle increases. The direction of slope is indicated by the azimuth of the radial line (isogonic line) on which the point appears. A horizontal slope is represented by a point at the plot center. A slope inclined 89° due east is represented by a point near the edge of the plot at the three o'clock position (i.e. on the N90°E isogonic line). Since slopes greater than 30° are uncommon, only the central part of the plotting sheet is needed. In the present study, the outermost circle of each diagram represents the loci of 30° slope angles. For convenience this circle is called the 30° isoclinal line. In addition the 10° and 20° isoclinal lines are shown.

The resulting plot is similar to the stereographic projection of crystal faces or the point plot used in preparing a petrofabric diagram. It should be noted, however, that normally radial coordinate paper is practical only where slope angles do not exceed about 25°. For the most accurate work the standard equal area projection net should be used (Chapman, 1952, p. 433). This type of net was actually employed in this study; the plot was made on the upper hemisphere in each case.

The density distribution of points over all parts of the plot is next determined and expressed by means of contour lines precisely as in a petrofabric diagram. The result is a statistical slope orientation diagram (SSO diagram).

The elements of the SSO diagram are analogous to those of the petrofabric diagram (Chapman, 1952, p. 438). A few of these need brief explanation. A concentration of points extending diametrically across the diagram and indicated by an elongation of the contour outline in that direction is known as a partial girdle. Such a pattern indicates a linear topography with valley-and-ridge trend perpendicular to the girdle length.

The clustering of slopes (points) between two relatively closely spaced concentric circles constitutes a small circle girdle. If such a girdle is discontinuous, it is called a partial small circle girdle. These girdles are roughly concentric with the projection (plot), and their presence generally indicates a high percentage of relatively steep slope and a low percentage of slope at a slightly lower angle of inclination.

The concentration of points within a small area is known as a maximum. A well-defined maximum is common near the plot center and represents a high percentage of gentle or horizontal slope.

For the detailed analysis we will be concerned with that part of the island shown in figure 2. Figure 3 represents a plot of 481 slope values obtained from that portion of the topographic map. To test the adequacy of sampling, two additional SSO diagrams, not included here, were prepared using alternate points on the grid, or 240 and 241 poles respectively. These two diagrams were similar in all major respects to figure 3, thus testifying to the validity of the sample and suggesting that one-half the number of measurements used in figure 3 would have produced a representative diagram.

The most obvious features displayed by figure 3 include a central maximum and a partial girdle trending roughly east-west. The central maximum indicates a high percentage of relatively flat ground, the partial girdle indicates a dominant set of roughly N-S ridges and valleys. The most frequent slope directions for slopes steeper than 5° are not diametrically opposed, but lie roughly along the N 90° E and S 65° W isogonic lines. This relation imparts triclinic symmetry to the diagram, and merely shows that one set of slopes strikes N-S and inclines eastward whereas the other set strikes N25°W and inclines WSW. The relation indicates a series of valleys with walls converging to the north, whereas intervening ridges taper to the south. This feature, brought out so clearly on the SSO diagram, is poorly shown on the topographic map. The average valley trend is about N15°W-S15°E.

The diagram indicates a noticeable absence of moderate and steep northerly and southerly slopes. For slope angle values above about 7°, easterly slopes are, on the whole, a few degrees steeper than westerly slopes. This asymmetry of ridge-and-valley cross section is poorly shown on the topographic map.

The maximum concentration (mode) is at the center of the diagram, and a comparatively large amount (about 20 percent) of the topographic surface is underlain by slopes with inclinations of less than 5°. The 6-12 percent con-

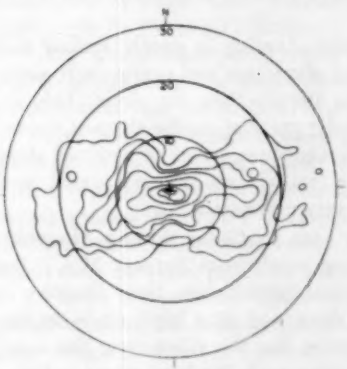

Fig. 3

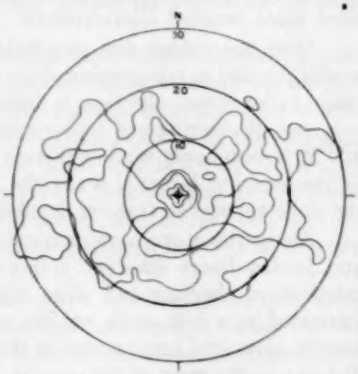

Fig. 4

Fig. 3. SSO diagram based on map data; 481 poles; contours 1-2-3-4-6-8-10-12% per 1% area of 30° circle.

Fig. 4. SSO diagram based on field data; 539 poles; contours 1-2-3-4-5% per 1% area of 30° circle.

tour lines surrounding the central maximum are conspicuously off-centered to the south and reveal a predominance of southerly slopes for these low angle values.

Figure 4 represents a plot of 539 slope values obtained by actual field measurements. Each measurement was made with a brunton compass and represents the slope for an area about 300 feet in diameter. These data were gathered as part of a detailed structural and petrologic study of the granite and before the concept of the SSO diagram was developed. Consequently, the method of sampling was not as systematic as it might have been. It was felt, however, that 539 readings should produce a representative diagram, since somewhat less than half that number is adequate when topographic map values are used. Figure 4, however, lacks continuity, and the 539 readings are apparently insufficient here where field measurements of slope are used. It is concluded that an SSO diagram prepared from topographic map data will show greater regularity and continuity in regional slope relations due to smoothing out of contours. A similar diagram prepared from field data tends to bring out the slight irregularities in slope, as well as the regional slope. A scattering of poles (points) and a large number of relatively weak maxima result. In spite of this, figures 3 and 4 show many striking similarities. Figure 4 shows the roughly E-W partial girdle, the central maximum, and the paucity of north slopes. The relatively high concentration of southerly slopes is probably due to the fact that a large number of field traverses were run along the southward trailing ridges. Had more E-W traverses (across the topographic grain) been run, a better diagram would have been obtained.

SHEETING

Sheeting is well developed over most of the hornblende granite area, and in general this structure roughly conforms with the present topography. The numerous departures from parallelism, however, are of great significance and need more detailed consideration.

Over the ridges and mountain summits sheeting is gently arched and nearly parallel to topographic slope. At lower elevations and particularly along steep valley sides. sheeting is inclined less steeply than the slope. This divergence of sheeting and topography is usually greatest on the steepest slopes, and it is commonly made apparent by a conspicuous series of inclined steps in the sheeting layers. Locally the step-like features are poorly formed, and the slope appears to transect the sheeting layers more smoothly.

Along the more asymmetrical ridges, such as Cedar Swamp Mountain and Jordan Ridge (the two ridges extending south from Sargent Mtn.), the relations of sheeting and slope are of special importance. Here sheeting is truncated at a low angle on the western slope and at a high angle on the eastern slope. So pronounced is this difference that the ridge crest lies some distance to the west of the crest in the sheeting arch. Such relations indicate that the ice tongue in the valley to the east cut back the eastern slope of the ridge nearly to its former crest. The ice in the valley to the west, however, failed to cut as deeply and steeply into the western flank of the mountain ridge. As a result asymmetrical ridges formed.

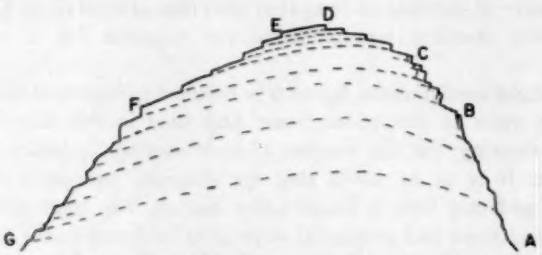

Fig. 5. Diagrammatic cross-profile of asymmetrical ridge, as viewed looking northward, showing relation of sheeting to topography.

Such a ridge is shown in figure 5. The steep western slope (F-G) shows the common relation of truncated sheeting and step structure. Steep eastern slopes (A-B) may show higher angle of truncation, but generally sheeting is difficult to observe here. Higher eastern slopes (B-C) are less steep and have been somewhat disintegrated by glacial ice and later frost action. Breaking down of such slopes appears to have been controlled largely by steep joints. On high western slopes (E-F) sheeting dominates over jointing in the control of slope angle. Gentler slopes, consequently, are encountered here. Near the crest (D-E) slope conforms with sheeting which is nearly flat. East of the crest somewhat steeper slopes (C-D) have formed by steps across nearly horizontal sheeting.

On the northwest slopes of the larger mountains, sheeting is nearly parallel to topography. In some places it dips a bit more steeply and in others a bit less steeply than the local slope. Some of the highest angles of sheeting truncation were observed on southeastern slopes where relatively gentle sheeting is exposed in the steep slopes. In the small hills known as The Bubbles (southeast of Eagle Lake), sheeting is nearly horizontal and is cut off at a high angle by the steep hillsides.

The above relations serve to indicate that most of the sheeting is preglacial in age. In a few places, however, it appears that a sheeting has formed parallel to the ice cut surfaces and is, therefore, glacial or post-glacial in age.

The relations between sheeting and topography, as well as the fact that sheeting surfaces appear closely spaced on ridges and more widely spaced further down the ice-cut slopes, favor the theory that sheeting in the hornblende granite developed in response to unloading, and is not controlled by the contacts of the granite body. This problem has been considered in some detail is an excellent paper by Jahns (1943).

Figure 6 represents a plot of 577 readings on sheeting. These data were gathered along the same series of unoriented traverses used in measuring the slope values involved in figure 4. Figure 6, however, shows excellent continuity and is considered to be a fairly accurate representation. This may appear somewhat surprising at first when it is realized that the data for figures 4 and 6 are not only essentially equally numerous but were obtained from the same stations. The reason for the greater continuity of figure 6, where the number of poles plotted is nearly the same, is due to the more uniform or more

regional nature of sheeting as compared with that of local slope. Consequently, notably fewer sheeting measurements are required for a representative diagram.

With slight modification, figure 6 is believed to represent the topography immediately prior to glaciation. Some data used in this diagram represent postglacial sheeting, but the number of such readings is believed to be relatively small. It is to be noted that the diagram represents only bedrock topography and very little is broad valley bottom. The greatest difference between sheeting slope and preglacial slope is to be found in the lower parts of the glacial valleys. Here the sheeting probably inclines a bit more gently than did the old preglacial slopes directly above it. This effect, however, is in part compensated for by the inclusion of some postglacial sheeting data. A truly representative diagram would probably show a slightly stronger girdle development with the outer contour not exceeding the 22-23° isoclinal line (circle) in the easterly and westerly directions.

The preglacial as compared with the present topography is considered to have been slightly less linear, because the east-west girdle of figure 6 is only weakly developed. This girdle, more accurately, trends about N80°W-S80°E and indicates an average stream line of N10°E-S10°W. The more nearly radial symmetry of figure 6 as compared with figure 3 indicates that easterly and westerly slopes were such more gentle prior to glaciation, whereas northerly and southerly slopes were slightly steeper.

In contrast to figure 3, figure 6 shows symmetry of easterly and westerly slope angles and a symmetry plane extending N10°E-S10°W across the diagram. Northerly and southerly slope angles, however, are asymmetrical. Extensive gentle slopes to the south and few gentle slopes to the north are shown by the off-centering of the 3-4-6-8-10 percent contours, and the somewhat greater abundance of steeper northerly slopes is shown by the slight off-centering of the 1-2 percent contours. This relation eliminates the possibility of an east-west symmetry plane and establishes a monoclinic symmetry for the diagram.

Raisz (1929, p. 141 and 155) felt that the topography immediately prior to glaciation was in the mature stage of development. Based on a comparison of SSO diagrams of other areas (Chapman, 1951), the writers conclude that the sheeting diagram (fig. 6) indicates an advanced stage of maturity for the Mount Desert Range prior to glaciation.

The differences between figure 6 on the one hand and figures 3 and 4 on the other, reflect general changes in topography produced by glaciation. The three most obvious changes appear to have been:

1. A shift in average trend of valley and ridge from N10°E-S10°W to N15°W-S15°E.

2. An increase in easterly and westerly slope areas and slope angles (in part at the expense of northerly and southerly slopes) due to the obliteration of earlier headwater areas by deeply carving tongues of glacial ice.

3. The development of an asymmetrical cross section for many valleys and ridges.

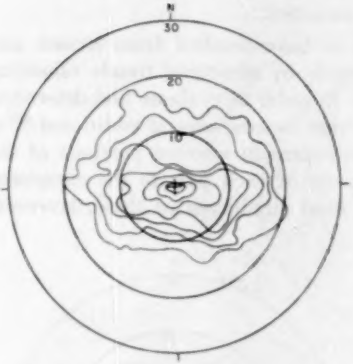

Fig. 6

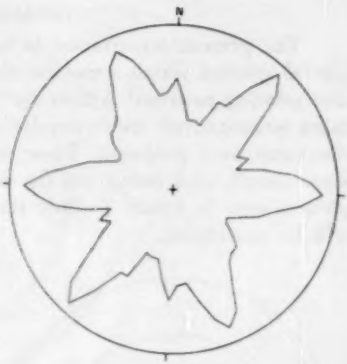

Fig. 7

Fig. 6. SSO diagram based on sheeting data; 577 poles; contours 1-2-3-4-6-8-10% per 1% area of 30° circle.

Fig. 7. Joint diagram based on 854 joint measurements. Radius of circle represents 40 joint measurements.

JOINTING

In addition to the studies of slope and sheeting, detailed observations on jointing were made in the granite area. About 3000 readings were taken, each representing the average attitude of roughly 20 nearly parallel joints. Joints were recorded as good, medium and poor; and the approximate spacing of each joint set was noted. These data were gathered at the same stations as those from which sheeting and slope data were obtained.

Jointing is a wide spread phenomena but is by no means uniform. Joints are sparse on steep valley slopes as compared with slopes at higher elevations. They are also less prevalent on broad rounded ridges than on the sharp narrow ones. The significance of these various relations can not be discussed here, but it will suffice to say that these studies indicate that jointing has in part controlled topography and topography in turn has had a control upon jointing. As between sheeting and topography, the interdependence of jointing and topography is convincingly demonstrated.

Figure 7 is a rose-diagram representing the strike of 854 separate joint measurements. Since most joints are steep to vertical, poles of joint planes were not used; but joint strikes, to the nearest 5°, were plotted.

Three distinct joint sets appear to make up the joint system of the area. These sets trend E-W, N40°E and N25°W. A more subordinate set trends N-S and destroys the otherwise well-developed orthorhombic symmetry pattern of the diagram. Comparing this diagram with figure 3, one notes a parallelism between the strike of westerly slopes and the N25°W joint set, and the strike of easterly slopes and the N-S joint set. In other words, the east valley walls parallel the N-S joints and west valley walls parallel N25°W joints. The significance of this relation will be considered later.

TOPOGRAPHIC ANALYSIS

The present topography is believed to have resulted from stream and glacial erosion which were controlled largely by structural trends (sheeting and jointing patterns) within the granite. In order to evaluate and determine more quantitatively the interrelations of these factors, several additional SSO diagrams were prepared. These diagrams represent selected portions of the slope profile and bring out the contribution of each part to the composite picture seen in figure 3. Only the three most significant of these, however, will be considered.

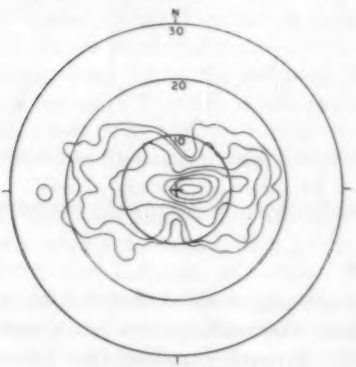

Fig. 8

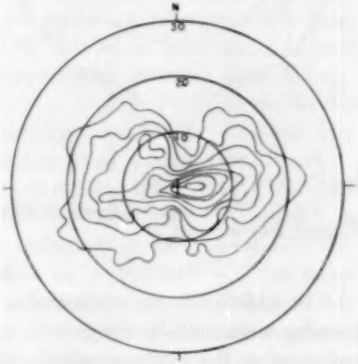

Fig. 9

Fig. 8. SSO diagram of highest slopes; 319 poles; contours 1-2-4-6-8-10-12% per 1% of 30° circle.

Fig. 9. SSO diagram of highest and next highest slopes; 565 poles; contours 1-2-3-4-5-6-7-8% per 1% area of 30° circle.

Figure 8 represents a plot of 319 slope poles obtained from the uppermost slopes of the highest ridges and mountain tops. As compared with figure 3 this diagram shows diametrically opposed slope direction concentrations, only moderate dispersion of slope direction, and marked asymmetry for gentle slopes (revealed by off-centered maximum).

Figure 9 represents a plot of 565 poles. It is a composite diagram and includes the poles of figure 8 in addition to those of the contiguous, slightly lower and steeper slope areas. In this diagram dispersion of slope direction is greater than in figures 3 and 8. An off-centered maximum for low slope angles is as pronounced as in figure 8. Like figure 3 but unlike figure 8, the suggestion of a slope direction concentration in the S65°W isogonic zone is apparent.

Figure 10 represents a plot of 718 poles obtained from the steepest slope areas which lie at elevations generally below those represented in figure 9. This diagram shows a lower dispersion of slope direction for the higher slope angles. The dominant slope directions, revealed by prominent maxima, are N85°E and S65°W.

Figures 8 and 9, in contrast to figures 3 and 4, show a strong maximum off-centered to the east about 2°. This striking difference appears to have the

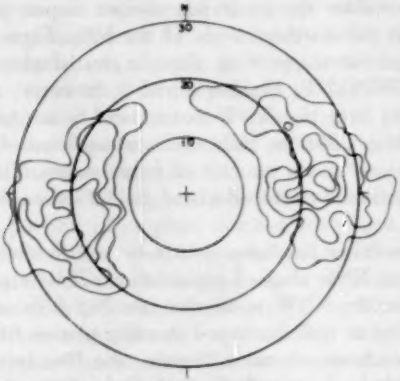

Fig. 10. SSO diagram of steepest slopes; 718 poles; contours 1-2-3-4-5% per 1% area of 30° circle.

following explanation. The gentle slopes (less than 10°) in figures 8 and 9 represent mountain and ridge tops only and reveal the asymmetry of glacial erosion at these high elevations. Corresponding slopes of figures 3 and 4, however, represent both mountain and ridge tops and valley floors; and the asymmetry at higher elevations is compensated by a reversed asymmetry at lower elevations. The asymmetry of mountain and ridge tops is, therefore, camouflaged in the more general type of diagram such as figures 3 and 4. The asymmetry of valley floors is due probably to deeper channelling by glacial ice in the western half of the valleys. In figure 3, which is based on map data, asymmetry of valley bottoms may be due in part to piling up of debris at the foot of east valley walls and the cleaning out of debris from the base of the west valley walls.

The effects of glaciation on the higher slopes, therefore, can best be determined by comparing figures 8 and 9 with figure 6 (sheeting).

Figure 8 shows an east-west girdle with slope direction concentrations nearly coincident with those for the gentler slopes of figure 6. This indicates there has been little or no change in the strike direction of the higher slopes due to ice action. The glacial effect appears to have been predominantly a rather uniform stripping essentially parallel to sheeting. The asymmetry of slope direction, as revealed by the off-centered maximum, however, shows that gentle easterly slopes have been increased greatly in area over gentle westerly slopes. This asymmetry appears to have been accomplished by high-level ice which moved southeasterly across the ridges and stripped the granite in layers parallel to sheeting. The stripping of any particular sheeting layer began at the free face of the steep U-shaped valley walls immediately to the east. Aided by joints, successive sheeting layers were stripped back toward the ridge crests to the west; and extensive stripping caused a westward migration of ridge crests and the consequent development of extensive areas of gentle slopes (C-D of fig. 5). The prominent N-S and N25°W joint sets facilitated this stripping and prevented a "rotation" of resulting ridge crests and newly cut slopes. The same maximum offset to the east is shown in figure 9.

Let us now consider the somewhat steeper slopes, generally at slightly lower elevations, at the northern ends of the hills. Figure 6 shows a rather uniform distribution for slopes with slope angles of about 7°-17° and slope directions of N45°W-N45°E. Figures 3 and 9, however, show a conspicuous bulging of contours into the NNW octant and a marked reentrant in the north direction. These relations indicate an abundance of NNW slopes, relatively few NNE slopes, and a paucity of north slopes. This shift in distribution of slopes is attributed to the effects of glacial ice on the north ends of the hills.

By comparison there has been relatively less flattening of NNW slopes than either north or NNE slopes by glaciation. This is probably due to the advance of ice from the NNW upon elongate N-S hills and the stripping of granite layers parallel to well developed sheeting planes. Stripping from NNW slopes may not have been extensive, because the free face from which stripping could be initiated was greatly limited. Only slight change in orientation of such slopes, therefore, could have resulted. The NE slopes, however, were differently affected. Lying very close to the steep freshly-cut west valley walls, where an extensive free face occurred, these slopes were subjected to extensive stripping which was undoubtedly greatly augmented by the N40°E and N25°W joint sets. This trimming back of slopes occurred in part at the expense of north slopes, and consequently north slopes were reduced in area. The significant change was perhaps the apparent clockwise rotation of generally northeasterly slopes; and the net effect was to greatly reduce the area of north slopes, to slightly reduce the area of NNE slopes, and to greatly increase the area of ENE slopes.

Let us next consider the effect of glacial action on the southern parts of these elongate hills. Figure 6 indicates a very uniform distribution for slope directions between S45°E and S45°W prior to glaciation. Figure 9, however, shows a distinct bulging of contours in the SSE octant indicating a relative increase in area and average inclination for SSE slopes. The diagram also shows a paucity of south slopes and a reduction in area of SSW slopes as compared with figure 6. These relations are essentially diametrically symmetrical with those of the northern counterparts.

The SSE slopes were most favorable for plucking action, because they lay on the lee side. The well developed sheeting, combined with the N40°E and N80°E joint sets, permitted extensive quarrying and the formation of stepped slopes which incline somewhat more steeply than the sheeting itself. The pronounced loss of south slope area is due in part to the cutting back of SSE slopes. Also responsible is the extensive slicing off of the southwest portions of the elongate hills. This action will be considered shortly.

The concentration of slopes in the N85°E isogonic zone of figures 6, 8, and 9 indicates little modification of slope direction for west valley walls by glacial action. The faint development of a slope concentration in the S70°W isogonic zone of figure 9, however, suggests considerable change in direction of east valley walls during glaciation. This is brought out more strikingly in figure 10 which represents only steep slopes. Figure 10 shows the strongest diametrical asymmetry with principal slope directions N85°E and S65°W.

Such relations indicate a well-developed linear valley pattern trending N15°W-S15°E with a conspicuous flaring of steep valley walls to the south. Incidentally this diagram shows also the greatest asymmetry in angle of slope with easterly slopes steeper than westerly slopes, and it brings out clearly the asymmetry of the ridge-and-valley cross profile. Figure 10, therefore, represents that portion of the slope profile which primarily determines the asymmetry of the general diagram (fig. 3).

The concentration of WSW slopes is believed to be due to the truncation of the southwestern portion of original north-south ridges. The cutting back of these southwestern slopes was accomplished by the valley ice which possessed a more easterly drift in the southeastern portions of the valleys. The easterly drift was due to a gradual mergence of the southward moving valley ice currents at low levels with the more southeasterly movement of the higher portions of the ice sheet. The faceting of these slopes was aided in large part by stripping along sheeting layers, but quarrying from the southern ends of the hills was in part a necessary prerequisite and accompanying mechanism. This quarrying action provided a free face which could be worked backward (in an up-ice direction) along sheeting planes. Stripping of successive sheeting layers was greatly facilitated by the N40°E and N80°E joint sets which helped to maintain a free face roughly perpendicular to ice flow.

The bevelling of sheeting layers on east valley walls was more pronounced in the southern parts of valleys where the easterly drift of ice was greatest. To the north, however, these surfaces gradually merge with the steep north-south slopes cut by the deeply eroding valley ice currents. An easterly drift of the upper ice is thus reflected in the modification of the east valley walls. Ice action on SW slopes, therefore, was similar but much more extensive and severe than on NE slopes.

The marked constriction of contour pattern along the N-S line of figures 3, 8, and 9, as compared with figure 6, is due in part to the reduction of north and south slope areas, as already described. It is also due in part to the destruction of considerable north and south slope area in the crestal region of the mountain range due to the formation of through-going glacial valleys. This dumbell-shaped contour pattern expresses not only the absolute loss of north and south slope areas but also the marked increase in areas of steep slope in the general easterly and westerly directions.

Figures 4, 8, 9, and 10 all show greater dispersion of slope direction for western slopes, between 10°-20°, than for corresponding eastern slopes. This difference does not appear from the topographic map to be due primarily to greater scalloping of westerly slopes. It seems more probable to indicate that eastern hillsides are straighter and more continuous, whereas western hillsides are arcuate in ground plan. Such asymmetry of outline supports the idea that abrasion and stripping by overriding ice were effective on the western half of hills, whereas, the eastern half was cut back intensely by valley ice currents.

SUMMARY AND CONCLUSIONS

The SSO diagram has proved to be an effective tool in the topographic study of this area. As shown by a comparison of figure 3 with three others

(not included here) of the same area but obtained by slightly different methods, independent plots yield diagrams similar in all major respects. In addition it has been shown that SSO diagrams reveal elements of topography not readily discernible from the topographic map. The picture presented by a topographic map alone is often misleading; unusual features, which may not be dominant features, often lead to poor interpretations of slope relationships. These diagrams simplify comparison of topography with such structural features as sheeting and jointing.

Since the area involved is small and the features relatively homogeneous throughout, the SSO diagrams should well represent this topographic unit. The use of an enlarged copy of the topographic map helped to reduce possible errors in slope measurement. It is known that reducing the number of poles by one-half did not alter the diagrams in many cases. Below a certain minimum number of poles, however, a diagram shows marked discontinuity and poorly portrays true slope relations. No method, other than experimentation, was attempted to determine this statistical minimum. It should be noted that SSO diagrams constructed from maps are no more accurate than the maps themselves. Although we might expect slope data obtained in the field to give the most reliable diagram, it has been shown in this study that many more field measurements are necessary to obtain a statistical sample, because of the high dispersion of slope values. It also appears that if somewhat random traverses are used in the collection of field data, they would best be run about transverse to topographic trends. Data on more regional features, such as sheeting, may be plotted from field measurements without a comparatively large number of readings.

Sheeting, which is shown to be controlled by the preglacial topography, is believed to have formed in response to unloading. A statistical plot of sheeting, therefore, is considered to represent rather closely the topographic character of the Mount Desert Range immediately prior to glaciation. A comparison of preglacial and postglacial topographies shows a conspicuous change in trend of valley and ridge, a marked increase in areas of easterly and westerly slopes, and the development of asymmetrical ridge-and-valley cross sections.

The trend of prominent ridges and valleys of this area is attributed to stream and glacial erosion and structural controls in the form of jointing and sheeting. The original valley sites were determined by preglacial streams whose trends were about N10°E-S10°W. Ice moving through these valleys actively scoured deep channels, due to concentrated flow, and removed the headwater areas of small streams in the crestal zone of the mountain range. Unusual long valley profiles thus resulted. Distinctly asymmetrical ridges and valleys were formed by quarrying activities of the ice as the greatly thickened glacier moved south-southeasterly across the mountain range. The coincidence of glacial and certain structural trends permitted extensive erosion of the southwestern portions of the elongate hills. This attack on southwesterly slopes was accomplished by a quarrying action along sheeting surfaces as the N25°W and N40°E joint sets formed block structures in the path of the overriding ice. Severe plucking occurred on SSE slopes to provide a free face for

more effective stripping on the southwesterly slopes. This stripping resulted in the development of extensive S65°W slope areas along the east valley walls. The overriding ice sheet was ineffective in modifying the trend of west valley walls, perhaps due to the control of N-S joints. There was, however, considerable reduction of slope inclination at higher levels. The asymmetry of valley-and-ridge cross profile was due in large part to abrasion along the stoss side and plucking along the lee side.

A marked linear topographic grain, trending N10°E-S10°W, had formed on rather homogeneous granite of the Mount Desert Range in pre-glacial time. The topography was bilaterally symmetrical with respect to this linear trend, and it may be considered to have possessed monoclinic symmetry. Pleistocene glaciation appears to have been directed toward the south-southeast at a marked angle to the N10°E-S10°W symmetry direction of the topography. So effective was this oblique attack by glacial ice that the monoclinic symmetry pattern of the old topography was supplanted by one of triclinic character.

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THE SYSTEM MgO—CO_2—A AND THE EFFECT OF INERT PRESSURE ON CERTAIN TYPES OF HYDROTHERMAL REACTION

R. I. HARKER

ABSTRACT. The study of polycomponent volatile systems hydrothermally raises difficult experimental problems. Thermal equilibrium data for the dissociation of magnesite, $MgCO_3 \rightleftarrows MgO + CO_2$, at varying pressures of CO_2 and inert gases indicate that the equilibrium temperature for this reaction is primarily dependent on the proportion of CO_2 at a given total pressure. In the region of a single metamorphic grade anomalously higher index minerals may be produced in carbonate-hydrate rocks, such as marls, than in the more purely carbonate or hydrate rocks.

INTRODUCTION

With the rapid development of high pressure-temperature techniques the knowledge of the chemical reactions involved in petrologic processes is greatly extended. Many parts of many systems have been studied under varying pressures of either water or carbon dioxide, which geologic evidence suggests are the most important volatiles in the earth's crust. However, in natural environments it is probably unusual for the volatiles present to consist of one component. This probability, together with several other possibilities (e.g., there may be more than one volatile phase, the system may not be closed, the rock pressure may not be the same as the pressure on the fluid phases) sometimes tends to be overlooked when the relatively simple experimental results of the laboratory are compared with field data.

The effect of these conditions is mentioned by Bowen (1940) with respect to the decarbonation of siliceous limestone and dolomite, and a recent plea for more experimental data on mixed volatiles was made by Thompson (1955). Danielsson (1950), MacDonald (1953), Thompson (1955), and Yoder (1955) discuss the situation in which the rock pressure is greater than a given hydrostatic pressure acting on the interstitial fluid. In this case the dehydration or decarbonation temperature of a reaction will be lower than it would if the rock pressure were equal to the given hydrostatic pressure on the fluid.

Yoder (1954) carried out an experiment with the dehydration of analcite in which the total pressure was appreciably greater than the water pressure. He found that 5,000 lbs/in.^{2} or 1,000 lbs/in.^{2} of water, at a total pressure of 30,000 lbs/in.^{2} (obtained by pumping the necessary amount of argon into the pressure vessel), gave the same results as 30,000 lbs/in.^{2} of water. Yoder suggested that this effect resulted either because the analcite structure did not break down on dehydration or that the argon proxied for H_2O.

EXPERIMENTATION AND RESULTS

In an effort to clarify the problem, data have been obtained on the dissociation of magnesite, $MgCO_3 \rightleftarrows MgO + CO_2$, where the total pressure was greater than the pressure of CO_2. The equilibrium temperatures for the reaction at varying total pressures of CO_2 were already known between 5,000 and 40,000 lbs/in.^{2} (Harker and Tuttle, 1955).

In the present study the same basic equipment was used and as before the equilibrium assemblages were determined by an optic and, when possible, x-ray examination of the solid phases in the products (op. cit.). Now, however, the hydrostatic pressure over and above that due to CO_2 was realized by pumping into the pressure vessel the desired amount of an inert gas.

In the first series of runs the pressure vessel, containing the magnesite, periclase and sometimes basic magnesium carbonate in separate crucibles as starting materials, was brought rapidly up to temperature, (in about ten minutes by appropriately prehating the furnace). The CO_2 was then pumped in until the desired CO_2 pressure was reached. This was then immediately followed by argon which was pumped until the desired total pressure was obtained. The results of these experiments are summarized by the series of P-T curves shown in figure 1. At initial CO_2 pressures of over 10,000 lbs/in.^{2} the temperature for the dissociation of magnesite depended only on the total pressure within observable limits. At a total pressure of 30,000 lbs/in.^{2} the result was indistinguishable from that obtained with 30,000 lbs/in.^{2} of CO_2. Thus

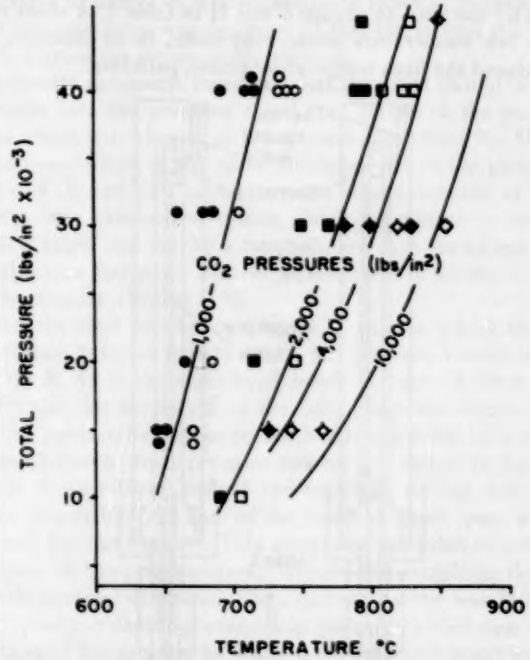

Fig. 1. Series of P_{Total}-T curves for the reaction $MgCO_3 \rightleftarrows MgO + CO_2$ at the following initial pressures of CO_2:

1,000 lbs/in.^{2}—circles,
2,000 lbs/in.^{2}—squares,
4,000 lbs/in.^{2}—circles on crosses,
10,000 lbs/in.^{2}—diamonds,

Black symbols indicate that $MgCO_3$ developed, white that MgO developed. These curves do *not* represent conditions of stable equilibrium.

far the results were parallel with Yoder's for the dehydration of analcite. At lower initial CO_2 pressures, however, a definite displacement of the P-T curve was produced. As can be seen from the figure a consistent series of results was produced.

Such experimentation assumes spontaneous mixing of the inert and the active gases. A useful interpretation of the data makes no assumptions as to the bulk composition of the vapor phase, nor even to the actual partial pressures of the different components of the vapor phase. Assumption is made, however, that the vapor phase in the pressure vessel is homogeneous. That this is *not* justified was proved experimentally in a second series of runs after this first series had been carried out.

In the second series a few experiments were made in order to find out whether the gases mixed spontaneously on injection one after the other into the pressure vessel. Data on the second series are summarized in table 1.

When helium was used as the inert gas it was found that the products of the runs depended on the duration of the run. The most extreme example of this is shown by the runs in groups I and II in table 1. A short run at 680°C produced the low temperature phase, magnesite, in all products. A long run at 660°C produced the high temperature phase, periclase.

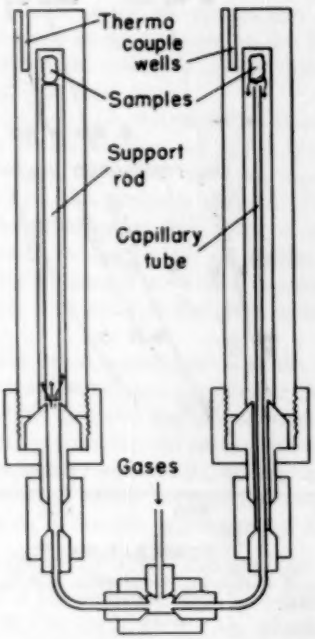

Fig. 2. Schematic diagram showing two pressure vessels linked to the pressure line in parallel. They are arranged so that the gases are injected at the bottom of the one on the left in the normal way but at the top of the one on the right (by way of the capillary extension through the basal cone).

In groups III and IV it is seen that by pumping the CO_2 in first and the inert gas second, magnesite is stable at a higher temperature (at least 826°C) than when the inert gas is pumped in first and the CO_2 second (less than 756°C).

For V and VI the apparatus was set up as in figure 2. The starting materials were supported as usual at the tops of the bores of the bombs but in one bomb the gases were injected at the top by means of a capillary extension running up from the cone at the bottom of the bomb, while in the other the gases were injected at the bottom in the usual way. As indicated in the table, the runs made in this pair of pressure vessels experienced the same total pressure, temperature, and duration; but where the gases were injected at the bottom magnesite was formed and where they were injected at the top periclase developed.

These runs strongly indicate that the gases were in fact not mixing fast enough to produce significant results. The gas to be pumped in first was concentrated in the region of the sample so that the partial pressure of the second gas was ineffective or nearly so. The exception to this was in the runs of group VI for which the apparatus was designed so that the second gas was made to displace the first from the region of the sample.

It was evidently necessary, therefore, that the gases should be mixed prior to their injection into the pressure vessel, i.e., mixed in the pump. For this series of runs argon was chosen as being more inert than N_2. Here we were faced with the possibilities of the gases unmixing (1) in the pump because of condensation of CO_2 and (2) in the pressure vessels because of the temperature gradient in the pressurized system. Such a gradient is bound to exist where there is a valve, and maybe a gauge, cooler than the sample holder, and the thermal diffusion factor for argon-CO_2 mixtures is known to increase appreciably with pressure (Becker, 1951).

The apparatus used for the next series of runs in which the gases were mixed in the pump prior to injection into the pressure vessels was set up as shown in figure 3. As in the right-hand bomb in figure 2 there was a capillary tubing running up the length of the bore inside the bomb. In this case, however, all the surrounding space in the bomb was filled by a baffle consisting of a piece of quarter-inch pressure tubing, not shown in figure 3, whose inner diameter fitted closely around the capillary tubing and whose outer diameter fitted closely into the bore of the bomb. A small space was left at the top of the bomb for the samples. This setup was intended to concentrate the bulk of the gases in the pressure vessel around the sample in the hot spot in order to inhibit thermal diffusion. The capillary tubing was led out directly through a "T" junction enabling samples of gas to be withdrawn directly from the hot spot around the samples as desired, see enlarged insert in figure 3 for detail. Such samples were then analyzed by means of a modified Orsat apparatus. The gases were collected over a saturated solution of sodium bicarbonate and then dried with concentrated sulphuric acid. Absorption of CO_2 was effected by using a 35 percent potassium hydroxide solution and argon was determined by difference. Samples of gas could also be withdrawn from the bottom of the pressure vessel and their compositions compared with those of

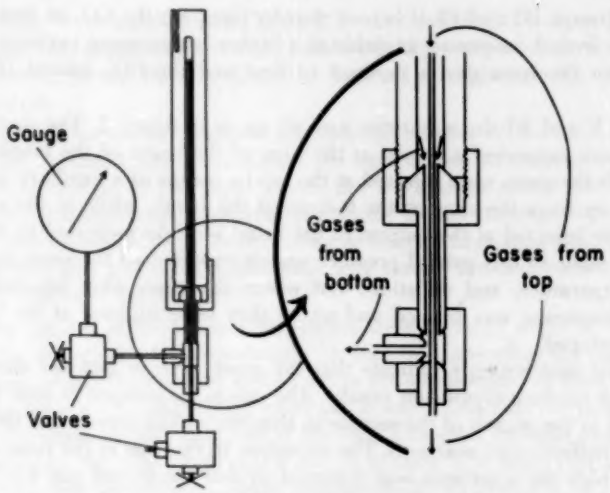

Fig. 3. Schematic diagram to illustrate the manner in which samples of gas could be withdrawn for analysis from different parts of the pressure vessel.

samples from the hot spot at the top. This was effected by closing the valve below the gauge and then removing the gauge. Through this same valve gases from the bottom of the pressure vessel could be led to the modified Orsat apparatus for analysis. No significant difference in composition was detected between the gas from the top and the gas from the bottom after a period of 120 hours with the hot spot at 510°C. This proved the effectiveness of the baffle in preventing thermal diffusion.

It was found by analyzing successive samples taken directly from the pump that a period of several hours had to elapse before the CO_2 and argon would form a homogeneous phase in the pump. This was so even at total pressures as low as 1,000 lbs/in.^{2}. An overnight waiting period was found desirable.

In this series of runs the gases in the pump were analyzed, after an adequate waiting period to insure homogeneity. The pressure in the bomb was then raised to the desired amount and the gas remaining in the pump at the newly elevated pressure was analyzed again. At the end of the run, a sample of gas from around the solid samples in the hot spot was withdrawn through the capillary tubing immediately prior to quenching. The starting materials normally used were periclase and either magnesite or basic magnesium carbonate. The quantity of material was kept to an absolute minimum, i.e., the smallest amount which could be studied optically. This was to reduce as far as possible the effect that either the absorption or liberation of CO_2 by or from the solids would have on the composition of the polycomponent gas phase delivered from the pump.

The results of the significant runs of this series have been summarized in table 2 and have been plotted in figure 4. The reasons for trace amounts of

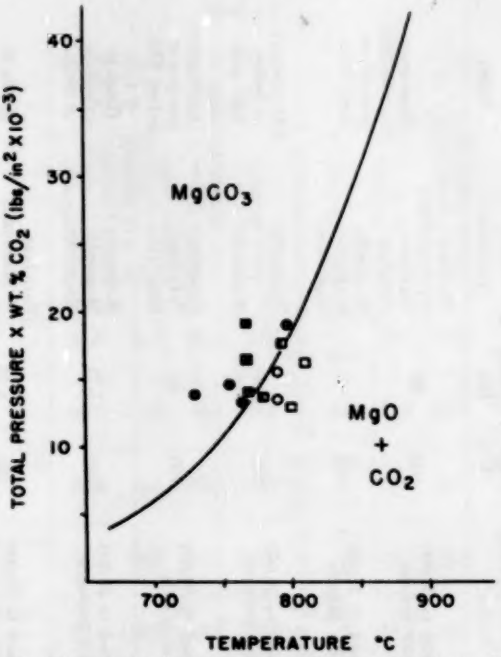

Fig. 4. P_{CO_2}-T diagram showing thermal equilibrium data obtained for the reaction $MgCO_3 \rightleftarrows MgO + CO_2$ with CO_2: A = ∞ (continuous line)
CO_2: A = approx. 1:1 by weight (circles)
CO_2: A = approx. 2:1 by weight (squares)
Black symbols indicate $MgCO_3$ is stable.
White symbols indicate $MgO + CO_2$ are stable together.

magnesite accompanying the periclase in runs 113 and 125 are not yet known. A powder photograph on a similar sample showed that the amount was too small for the very strong {10*4} reflection to be seen at all. As the periclase in runs 112 and 124 was free from any visible magnesite it was considered that these runs were made outside the stability field of magnesite. The P-T curve shown in figure 4 is that which was determined for the same reaction but with 100 percent CO_2 (Harker and Tuttle, 1955). It can be seen that even with argon-CO_2 mixtures those runs in which magnesite was the stable reaction product lie on the low temperature side of this P_{CO_2}-T curve and those in which periclase was produced in the products lie on the high temperature side. It therefore appears that the effective pressure of the CO_2 in the mixed volatile runs is directly proportional to the weight percentage of CO_2 in the gas phase.

CONCLUSIONS

These data show that at CO_2:A ratios of approximately 1:1 and 2:1 by weight, and at the pressures and temperatures studied, the equilibrium dissociation temperature of magnesite is dependent on P_{CO_2} rather than P_{Total} and

TABLE 1

Group	Run	Starting Materials	1st Gas pressure lbs/in.²	Total Gas pressure lbs/in.²	Temp °C	Time Hrs	Solid Products	Notes
I	1383	Magnesite	1,000 CO_2	Raised to 31,000 with helium	660	68	Periclase (+magnesite)	Time more effective than pressure in these runs.
	1384	Basic Mag. carb.					Periclase	
	1385	Periclase					Periclase	
II	1389	Magnesite	1,000 CO_2	Raised to 31,000 with helium	680	6	Magnesite	
	1390	Basic Mag. carb.					Magnesite	
	1391	Periclase					Magnesite	
III	1191	Periclase	10,000 CO_2	Raised to 30,000 with nitrogen	826	5	Magnesite	CO_2 pumped in before inert gas.
	1192	Magnesite					Magnesite	
IV	1407	Magnesite	20,000 N_2	Raised to 30,600 with CO_2	756	53	Periclase	Inert gas pumped in before CO_2
	1408	Basic Mag. carb.					Periclase	
	1409	Periclase					Periclase	
V	1413	Magnesite	1,000 CO_2	Raised to 20,000 with nitrogen	680	5.75	Magnesite	Gases in at bottom as usual.
	1414	Basic Mag. carb.					Magnesite	
	1415	Periclase					Magnesite	
VI	1416	Magnesite	1,000 CO_2	Raised to 20,000 with nitrogen	680	5.75	Periclase + Magnesite	Gases in at top.
	1417	Basic Mag. carb.					Periclase	
	1418	Periclase					Periclase	

Brackets show that only a small amonut of the phase indicated appeared in the products.

TABLE 2

Run	Starting Materials	Vol. % Argon at 28°C in pump: After mixing, before pumping to high pressure.	Vol. % Argon at 28°C in pump: At high pressure after pumping up pressure vessel.	Vol. % Argon at 28°C in gas from samples in bomb	Time in hours	Temperature in °C	Total Pressure in lbs/in.^{2} x 10^{-3}	Total Pressure x Vol. % CO_2 (28°C) in lbs/in.^{2} x 10^{-3}	Total Pressure x Wt. % CO_2 in lbs/in.^{2} x 10^{-3}	Solid Products
110	Basic Mag. carb.	43.8	44.7	44.5	42.5	797 to 800	33.1 to 33.5	18.5	19.2	Magnesite
111	Periclase									Magnesite
112	Magnesite	50.7	50.2	50.2	89.0	788 to 789	26.0	13.0	13.6	Periclase
113	Periclase									Periclase ((+magnesite))
124	Basic Mag. carb.	51.6	49.5	50.5	40.0	787 to 791	30.0 to 30.5	15.0	15.7	Periclase
125	Periclase									Periclase ((+magnesite))
126	Basic Mag. Carb.	52.5	52.0	53.3	100.0	727 to 730	27.5 to 28.8	13.2	13.8	Magnesite
127	Periclase									Magnesite
135	Basic Mag. carb.	52.5	52.5	52.0	43.0	750 to 758	29.4 to 29.6	14.2	14.9	Magnesite
136	Periclase									Magnesite
143	Basic Mag. Carb.	53.9	54.0	54.1	90.0	761 to 765	26.0 to 29.2	12.6	13.3	Periclase (+magnesite)
144	Periclase									Magnesite
147	Magnesite	40.4	40.3	40.5	42.5	788 to 796	28.8 to 29.5	17.3	18.0	Periclase + Magnesite
148	Periclase									Periclase + Magnesite
152	Magnesite	36.5	36.4	36.4	96.0	764 to 771	29.2 to 29.4	18.6	19.3	Magnesite
153	Periclase									Magnesite
156	Magnesite	36.2	37.2	37.2	120.0	777 to 783	20.5 to 21.4	13.2	13.6	Periclase (+magnesite)
157	Periclase									Periclase (+magnesite)
163	Magnesite	38.8	38.7	38.8	120.0	771 to 775	21.8 to 21.9	13.4	13.9	Periclase (+magnesite)
164	Periclase									Periclase (+magnesite)
186	Magnesite	36.4	35.7	36.9	99.0	788 to 800	19.6 to 19.8	12.4	12.9	Periclase
187	Periclase									Periclase
193	Magnesite	33.7	34.4	34.2	69.5	809 to 811	23.6 to 24.0	15.7	16.2	Periclase
194	Periclase									Periclase
195	Magnesite	36.3	37.3	37.7	98.5	765 to 775	25.4 to 25.8	16.0	16.5	Magnesite
196	Periclase									Magnesite

Brackets show that only a small amount of the phase indicated appeared in the products.

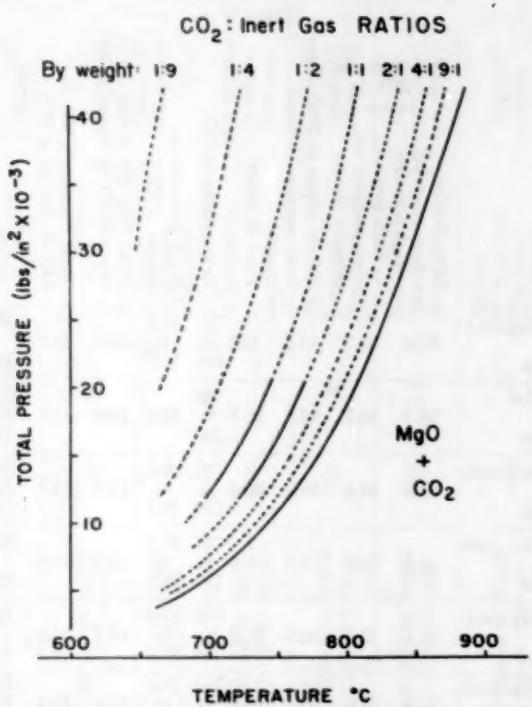

Fig. 5. P_{Total}-T diagram showing the thermal equilibrium curves for the reaction $MgCO_3 \rightleftarrows MgO + CO_2$ as deduced for varying proportions of CO_2 to inert pressure. The continuous lines are the regions in which experimental results confirm the curves.

can be approximately estimated from the P_{CO_2}-T data previously published. Assuming that the same dependence on CO_2 pressure holds for other ratios of CO_2 to inert material, it is possible to draw a series of P_{Total}-T curves for different proportions of CO_2 in the volatile phase (see fig. 5).

It is not unreasonable to suppose that this relationship is generally and approximately true at least where the compressibilities of the solid phases are insignificant relative to the compressibilities of the volatiles. The results with CO_2-inert gas mixtures are in accord with what was logically anticipated from Roy's (1954) theoretical consideration of water systems, and the reader is referred especially to his lucrative cubical P_{Total}, $\frac{P_{H_2O}}{P_{Total}}$, T diagram, (p. 136, op. cit.).

When correlating and comparing experimental with petrologic data, therefore, the proportions of the volatile components should be considered in addition to the total pressure (overburden). To take a specific example, consider a calcite-quartz assemblage in a dry massive siliceous limestone at an estimated depth of approximately 5 miles. If after a period of contact metamorphism at a pressure of about 2,000 atmospheres such an assemblage survives with little change except recrystallization, then it is highly probable

that the temperature did not exceed 800°C. If the massive limestone is now replaced by a marl (containing readily dehydrated minerals) the above deduction can be made with even more certainty. The CO_2 pressure in this second case must be appreciably less than the total pressure of water and CO_2 together. This means that in different rocks the upper limit of thermal stability of the calcite-quartz assemblage is not fixed by total pressure and temperature. Under a given set of physical conditions a decarbonation reaction will proceed more readily in an impure wet limestone than in a pure dry one.

Obviously the effect of CO_2 and/or carbonate impurity in a shale will affect the dehydration reactions in an analagous way. One must not be surprised therefore to find that in a metamorphosed sedimentary series the marls and calcareous shales are metamorphosed to an apparently higher grade than the associated more purely hydrous or carbonate rocks.

The value to petrology of experimentally determined P_{CO_2}-T and P_{H_2O}-T curves is not jeopardized by the present study. The points along the curves still represent the *maximum temperatures at which certain phases can exist or coexist.* The extra variable which must be taken into account when considering reactions involving volatiles means that more experimental data are needed, especially on those reactions which involve only solid phases and are insentitive to pressure or only slightly sensitive to the *total* pressure. Such reactions as Wollastonite + Monticellite $\rightleftarrows$ Akermanite (Harker and Tuttle, 1956) and the solid solution and exsolution of carbonates and of feldspars should prove especially useful.

Finally it may be concluded that the collection of meaningful data on mixed volatile systems is not easy. Diffusion coefficients and thermal diffusion factors cannot be ignored at elevated pressures and every care is essential to insure a homogeneous gas phase of known composition throughout the experiments.

Perhaps the most satisfactory way, when possible, is to weigh the volatile components into either a sealed tube or a Morey bomb as was done by Roy, Roy, and Osborn (1953). With a Morey bomb it would be very adviseable to use the finest possible capillary tubing to link the bomb to the pressure gauge in order to discourage any differential condensation in the gauge away from the hot part of the system. In addition, the bomb itself could be rocked in the furnace to promote homogeneity in the volatile mixture around the samples.

The sealing of weighed volatiles such as CO_2 and water into small platinum tubes presents a few problems but it can be done. No unusual difficulties are encountered with volatiles which at room temperature are combined as solids and liquids such as ammonium salts and inorganic acids. In using sealed tubes, however, it must be borne in mind that if a solid phase is to react with only one of the components in the vapor phase, or with one component at a time, then unless the amount of solid is very small indeed such reaction may upset the composition of the vapor phase to an appreciable extent.

ACKNOWLEDGMENTS

Part of this work was carried out at the Pennsylvania State University with the financial support of the National Science Foundation. The author is indebted to the helpful discussions with, and advice received from, Drs. O. F. Tuttle, R. Roy, J. W. Greig, and A. B. Cummins.

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ESSAY REVIEW

Mechanik der Erde; by RICHARD A. SONDER. P. viii, 291; 91 figs. Stuttgart, 1956 (E. Schweizerbartsche Verlagsbuchhandlung, DM 42.00).—Dr. Sonder of Olten, Switzerland, presents a unified tectonic theory designed to account for earth behavior as revealed by geological and geophysical observations. To this overwhelming task, he brings a background in mechanics, a wide acquaintance with geologic literature, a flair for quantitative treatment, and the boldness (in some cases rashness) to write equations and to enter values in any field of geology, where in many cases the specialists have feared to tread.

The author's style is clear and direct, and the organization of subject matter is excellent. Particularly commendable are the tables and graphs, some of them fold-outs on heavy paper.

Mechanik der Erde consists of one section dealing with tectonic processes as revealed by geology, and another devoted to their interpretation on the basis of the author's geomechanical theory. This arrangement is not altogether satisfactory, for the presentation of the first part lies so much under the spell of the theory, dealt with in the second, that the reader who begins with part 1 may fail to follow some of the author's lines of thought.

Sonder assumes that the earth consists of a central core, a mantle, lacking in strength, and an outer crust or tectonosphere. As the core contracts, the weight of the outer layers causes tangential compressive stresses. The mantle fails by plastic flow, but the stronger crust is assumed to behave elastically, according to Hooke's law, even over long periods of geologic time. At the beginning of a typical "geocycle" the crust has a minimal tangential stress. Ocean basins are deep, continents are high, and glacial conditions are likely to arise (hence this stage is termed "cryogenic"). As the core shrinks in the advancing cycle, the crust is tangentially compressed, like a giant spring. But, whereas the stresses are uniformly distributed over the globe, the resulting elastic strains are not; there is experimental evidence that granitic rocks are more compressible than are gabbroic ones. Accordingly, the continental segments undergo a greater compression, and a greater increase in density, than do the oceanic ones; continental segments will as a consequence subside isostatically, relative to the ocean basins, by amounts of tens to hundreds of feet. Transgressions result, and the complicating factors of erosional unloading and sedimentary loading enter the picture.

When the tangential stresses within the crust exceed its strength, then failure occurs abruptly. This is likely to be localized in geosynclinal regions, where the strongest (middle) members of the crust are relatively thin, and where steep downwarping of the crust favors shear failure. The elastic strain stored up in the vicinity is released in tangential expansion resulting in thrusting, folding and flowage of crustal material along the break—orogeny. This expansion reduces the density of the region involved, and results in isostatic rebound.

In the typical geocycle, local orogenies occur sporadically as the stresses mount, but the stresses and elastic strains within the crust as a whole continue to build up until they can overcome the strength of the crust along great-circle routes—the great mobile belts of the earth—a process believed to require a

time-span of around 200 million years. When such failure finally occurs, the stresses stored within the entire crust are relieved by a series of orogenies (orogenic state). The last three geocycles are the Caledonian, the Variscian, and the Ando-alpine; we are living in the beginning stages of a new one. Many other fascinating ramifications of the theory cannot be dealt with here.

The first part of the book, dealing with geological considerations, includes discussions of sedimentation, orogeny, epeirogeny, rhegmageny (transcurrent or "wrench" tectonics), taphrogeny, phorogeny (displacements of the crust relative to the core), pyrogeny (volcanism), and crustal structure. The author admits that, in view of space limitations, this is essentially a statement of his own views and predilections; the reader will find many of them to be, mildly stated, controversial. Geologists with some background in sedimentology will be jolted by simple tectonic interpretations of rock types, bedding, unconformities etc. Few will care to follow interpretation of lithic alternations such as 15,000 graded beds in the alpine Flysch as a record of so many tectonically induced sea-level oscillations, or will accept accordant levels of Guyot tops as evidence of six Precambrian geocycles. The cheerful use of "average" rates of sedimentation and — piling inference on inference — of "average hiatus factors" will meet with considerable misgivings. But though the reader may at times tear his hair at the author's tendency to be carried away by his passion for numbers at all cost and for simple mechanical explanations, he will find the work as a whole to be an astonishingly imaginative and well integrated synthesis.

The theory is, above all, remarkable in providing a mechanism which links orogeny and epeirogeny, and which affords some explanation for major features of earth history; to pick one detail, it is highly attractive as an explanation for cyclic sedimentation, essentially fulfilling the requirements of M. Weller's diastrophic control theory. While many problems such as the origin of island arcs on oceanic crust, the widespread development of tensional faulting, and the large role which extrusive volcanism has played in earth history are not integrated into the theory in its present form, they could probably be reconciled with it by minor modifications.

Several theoretical obstacles may be raised. Some will question the basic premise that the crust will accumulate and retain elastic tangential strain in accordance with Hooke's law. Those who believe that the crust would react to long continued pressures not by elastic compression but by plastic flow, will *ipso facto* be obliged to reject the core of the theory. Another difficulty, raised by the author himself, concerns the amount of core shrinkage envisioned. Sonder believes that a shrinkage of at least 1 percent of the earth's radius per geocycle (200 million years) is needed to account for the geological effects. This shrinkage is far in excess of any which he can account for by cooling, hence he must appeal to profound changes in the state of matter within the core.

This is an extraordinary book. As a challenge to tectonic thought it has few equals; as a catalyst, it is bound to generate a vigorous reaction.

A. G. FISCHER

REVIEWS

Sedimentary Rocks, 2d ed.; by F. J. PETTIJOHN. Pp. xvi, 718; 173 figs.; 119 tables; 40 pls. New York, 1957 (Harper & Brothers, $12.00).—The second edition of Professor Pettijohn's successful book is greatly expanded, re-worked, and brought up to date. Many new figures have been added and, though the number of plates is the same as in the first edition, much new material has been added by economizing space. The same superb quality of the plates is maintained in the new edition; plate legends have been re-organized, placing the origin of the specimen first. A very useful new feature is the inclusion of references in working bibliographies at the end of appropriate sections of the text, instead of in footnotes, as in the first edition. The authors and subjects are indexed, but the plates, tables, and figures are not; such a list is sorely missed by the reviewer. The language of the text has been improved in many instances, though numerous ambiguous attempts at definitions and repetitions of different terms for the same thing mar some chapters; errors are few.

A vast field of knowledge, represented by an enormous literature, has been brought together by Professor Pettijohn, who is probably the best qualified person to do this. The reviewer appplauds the author's emphasis on measurement of oriented structures made by currents (called "paleocurrents" by Pettijohn), but raises the same objection against Pettijohn's use of the term *vector* to describe an orientation that McIntyre and Christie (Geol. Soc. America Bull., v. 68, no. 5, p. 648) cite in discussion of a paper by J. H. Hodgson (*ibid*, p. 616-643). Other features that the reviewer dislikes are the use of old-fashioned language of oxides in discussing chemical analyses, the mistaken substitution of *lamination* for lamina throughout the book, and the confusion between "clay" and "shale" present in Chapter 8 (Shales, Argillites, and Siltstones).

Pettijohn accepts the importance of turbidity currents (which he invariable calls "turbidity flows" or "underflows") in the resedimentation of coarse arenite in deep water, but gives the impression that they are "abnormal" by contrasting them repeatedly with "normal" currents. It should be emphasized that not all turbidity-current deposits are graywackes (Pettijohn notes this by mentioning graded limestones), nor are they all arenites with muddy matrix (framework disrupted). Recent deep-sea sands are "clean" (framework intact). Pettijohn's classification emphasizing fluidity (density and viscosity) of depositing agent (as inferred from presence or absence of muddy matrix), therefore, fails to define unique categories, for Kuenen has shown that even a turbidity current of high density possesses low viscosity and therefore differs little in its physical parameters from a stream in flood; the same flow formulae were found to apply to each. The sedimentary structures in the deposits laid down by turbidity currents and streams do differ, however, as has been well shown, beginning with Bailey's recognition of two contrasting sandstone types; the difference is amply documented by Pettijohn.

Those who enjoyed Pettijohn's first edition will like the second even more; all students of sedimentary rocks will find the second edition a necessary reference, both as an introduction to the field and source of further reading. The same general arrangement of topics (structures and textures, kinds

of sedimentary rocks, dispersal and larger association) is retained with slight modification but with considerable change in emphasis. The chapter of the first edition entitled "Weathering" has been replaced by one on "Provenance (and Mineral Stability)," "Transportation" by "Dispersal," "Deposition" by "Depositional Environments" and a new final chapter, "Historical Geology of Sediments" added. Most stratigraphic concepts have been eliminated.

The reviewer applauds both author and publisher for the new edition, which stands in a class by itself. JOHN E. SANDERS

Chazyan and Related Brachiopods; by G. ARTHUR COOPER. P. 1245; 268 pls. Washington, D. C., 1956 (Smithsonian Miscellaneous Collections, vol. 127, in two parts, $20.00).—A few works are truly epochal, in the sense that they change standards, and make obsolete what had been acceptable procedures. This is such a book. True, it has had harbingers, the joint monographs with Schuchert and with Ulrich, but here for the first time we have Cooper's methods and concepts consistently applied to the detailed description of a large and varied fauna. Study of Ordovician brachiopods will never be the same again.

The title describes the planned scope of the study, but more and more ancillary material has been added until in its final form this book is rather a monograph of the lower middle Ordovician brachiopods of North America.

An extremely useful introductory section provides a stratigraphic setting, with description of all the formations in which brachiopods have been found, and suggestions on the correlation and relationships of these units. For a few areas this material is drawn from published records, but for most it reflects the author's personal knowledge of the rocks, and is correspondingly valuable. The interval between the Canadian and the Trentonian is divided into five stages, proposed jointly with Byron N. Cooper. Acceptance of these stages will depend, of course, on the extent to which workers in other groups find them useful. Personally, I welcome the Wilderness stage, which includes Black River and Rockland of the standard formation scale. In recent years it has become more and more apparent that there is no faunal break below the Rockland, but a major break above it. The new stage lets us recognize this reality, and leaves the Trentonian as a homogeneous unit for post-Rockland beds.

There is little need for comment on the brachiopod descriptions themselves. The validity of the hundreds of new species and scores of new genera will be tested as others try to identify their fossils in terms of Cooper's taxa. But in general it may be said that the descriptions are full and clear, and the illustrations admirable, especially in the extent to which variation within the species is indicated by a number of photographs. A fuller treatment of previously described genera would have been welcomed, but was no doubt omitted to save space. A few general remarks may be in order.

It is to me a treat to read a brachiopod description in which I am sure which valve is being discussed without turning back to the section on morphology to see what *this* author means by dorsal. Perhaps specialists in the group do not appreciate how confusing to others is the use of the terms dorsal and ventral. Brachial and pedicle may be bad words from many points of

view, but they have the great virtues of clarity and consistency of usage, which may well outweigh many defects.

If there are any not yet convinced of the value of acetic acid etching as a preparation technique, let them study the amazing series of phosphatic shells Cooper has been able to extract by its use.

It may be noted, with appreciation, that Cooper, in erecting 69 new genera, has found 69 names which are fairly short, appropriate, pronouncable and esthetically pleasing. Such things do not happen by accident, and we must be grateful for the care with which these names have been coined. Let those who would excuse cacaphonies with the plea that all the good names have been used, take note.

G. WINSTON SINCLAIR*

* Published with permission of the Deputy Minister, Department of Mines and Technical Surveys, Ottawa.

Ordovician Trilobites of Argentina; by HORACIO J. HARRINGTON and ARMANDO F. LEANZA. P. 276; 140 figs. Lawrence, 1957 (Dept. of Geol., Univ. Kansas Special Pub. No. 1, Univ. Kansas Press, $6.00).—This volume will prove indispensable not only to students of trilobites but also to anyone interested in Ordovician stratigraphy in South America. Because an increasing number of species and/or genera known previously only in South America are being found in North America, this study will be particularly important to people working on Ordovician rocks in the Great Basin and in the Marathon Region of Texas.

The authors have divided the paper into two main sections, the first dealing with stratigraphy, the second with taxonomy. All the important Ordovician sections are located and described in detail within each of the three main areas of Ordovician outcrop. Vertical distribution of fossils within these sections leads to the establishment of 18 faunal zones, some based on trilobites and some on graptolites.

Harrington and Leanza close the section on stratigraphy with a history of Ordovician sedimentation in Argentina. From the Tremadoc deposition to the Arenig-Llanvirn deposition there is a nearly complete reversal in the east-west facies relationship. In both the Tremadoc and Llanvirn-Caradoc beds there are coarse deposits which are attributed to glacial origin. Students of the Johns Valley shale of Oklahoma may find a familiar ring in the description of "exotic" blocks of San Juan limestone found within the Rinconada formation, both of Ordovician age.

Readers may find the use of the prefixes unda-, clino-, and fondo- unfamiliar and they may wish to go back to Rich's (1951) original paper to refresh their memories on the use of these terms. (They literally mean "wave" (Latin), "to slope" (Greek), and "bottom" (Spanish).) "Shallow", "medium", and "deep" would convey the same practical meanings.

The taxonomic portion of the paper deals with 124 species (70 new), 83 genera (26 new), and 2 new families. Here are assembled descriptions and excellent illustrations of species which have previously been scattered through various German, Japanese, and Argentine journals. Descriptions and synonymies are brought up to date, untangling several knotty taxonomic problems. Experience had begun to show that this section would be important to North

America workers long before the book was published, particularly in the identification of olenid trilobites. I believe that the new genus *Keidelia* may prove to be more closely related to *Carolinites* and Komaspids than it is to proetids, but this is the only even mildly adverse comment I have mustered on the taxonomy. (Professor Harrington, written communication, June 9, 1957, confirms my belief.)

The Department of Geology and the Press at the University of Kansas are to be congratulated on this first Special Publication. The format is excellent and the half tone illustrations equal in quality to many collotype reproductions. The text and figures are placed close to one another for easy reference.

REUBEN J. ROSS, JR.*

* Publication authorized by Director, U. S. Geological Survey.

General Geography For Colleges; by O. D. VON ENGELN and BRUCE C. NETSCHERT. P. xiii, 681. New York, 1957 (Harper & Brothers, $7.50).—Messrs. Von Engeln and Netschert have without a doubt achieved their aim "to promote geographic literacy" in this recent text. The first one-sixth of the book covers the physical fundamentals of geography: the major world landmarks such as landform regions, water bodies, cities, as well as elements of cartography and climate. This is followed by a rather detailed regional geography of the world based upon climatic regions.

The use of climatic regions as a framework for regional presentation is a time-honored and time-tested technique in geography. Although this approach is not to every instructor's liking, it is nevertheless capably presented in *General Geography For Colleges.* Two inevitable problems arise, large countries such as the United States and the Soviet Union with more than one climatic type are described in several chapters. Likewise, the relationship of world industrial centers to climatic types is often much less direct than the relationship of agricultural activities and climate. This technique does, however, constantly emphasize the comparison between world regions having the same climate. Eleven climatic types are described and causes are analyzed, after which all the world regions with a certain climatic type are presented. The resulting regional coverage often is equal in factual material to that found in some regional texts.

Several facts about the presentation of material in this text should be mentioned. Maps are mostly half-tone with relief shading, and with one or two exceptions are very legible. Climatic boundaries and major economic activities are shown in white; the more important an economic activity, the larger the lettering. Maps thus more or less present a visual summary of the text material. Inset maps of various foreign climatic regions superimposed upon a map of the United States simplify size comparisons with the United States. Students will no doubt appreciate the fact that key words in the text are in boldface type. Many excellent aerial photographs further enhance the reading. There is some doubt as to the desirability of including a predominantly British suggested reading list in the appendix since many of these books might not be available in American libraries.

ERICH F. BORDNE

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